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## CORRECTIONS AND EMENDATIONS

Contributors to Volume 53 have been invited to send corrections and emendations to be made in their papers, and the volume has been scanned with some care. Corrections and insertions are as follows:

Page 348, line 14, *for* Fig. 2, *read* Pl. 1.
" 361, line 7, *for* vest, *read* vent.
" 456, line 14, quotation *should read* "relatively unaltered limestone".
" 468, line 7, *for* with, *read* within.
" 489, line 6, *for* Plate 5, *read* Plate 4.
" 538, Plate 1. Lines indicating a hiatus should be drawn in the space between the words "Williamson shale" and "Wolcott limestone" in column headed "Western New York, G. H. Chadwick, E. R. Cumings, G. M. Ehlers."
" 612, Plate 3 B. A narrow belt of Sherman Fall along the south shore of Lake Clear north of the printed "Fault" has been omitted.
" 646, Plate 7. Areas of Hull limestone on southern Wolfe Island and south of Cape Vincent have been mapped with the Rockland limestone.
" 647, Plate 1. Granitic dike on Cass House Cr. passes under thrust sheet, does not cut it as shown.
" 722, Figure 7 is upside down.

Periodical

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- Correlation of the Silurian Formations of North America. By Charles K. Swartz, Chairman.

AIR VIEW LOOKING EASTWARD ACROSS THE HEAD OF KIPAHULU VALLEY INTO THE SO-CALLED HALEAKALA CRATER

The steep cliff on the left is the west wall of Kaupo Valley, and the dim cliff on the right is the west wall of Keanae Valley.
Photo by U. S. Army Air Corps.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 1-14, 1 PL., 3 FIGS. JANUARY 1, 1942

# ORIGIN OF HALEAKALA CRATER, ISLAND OF MAUI, HAWAII

BY HAROLD T. STEARNS

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## ABSTRACT

A geologic map of the summit depression of Haleakala, Maui Island, Hawaii, was finished by the writer in 1935 as part of the systematic geological investigation of the Hawaiian Islands by the United States Geological Survey, in co-operation with the Territory of Hawaii. The depression is 7 miles long and 2 miles wide. The highest point on the rim is 10,025 feet above sea level. Evidence is presented to support Cross' theory that this "crater" is due chiefly, and perhaps entirely, to stream erosion by the recession of two great amphitheater-headed valleys, Keanae and Kaupo, from opposite sides of the mountain, instead of being due to either collapse, sliding away of a side of the cone, or explosion, as set forth by others. Kipahulu, Waihoi, and Manawainui valleys were likewise formed during this long erosion period. The ancient main rift zone crossed the summit and the heads of Keanae and Kaupo valleys. Renewed volcanic eruptions along this rift built large cinder cones and poured out voluminous flows that nearly masked the ancient divide between the two valleys and partly filled both to the sea. The depth of the fill



Photo by U. S. Army Air Corps.

probably exceeds 2,000 feet at the valley heads. The lava flows displaced the drainage channels and caused the streams to undercut their valley walls, thereby widening their heads. Many of the late eruptions took place along fissures crossing the valley walls, thereby giving vent to flows that cascaded down the steep walls. A broad thick mud flow underlies the late lavas at the mouth of Kaupo Valley.

## LOCATION AND AREA

"Haleakala Crater," in the Haleakala Section of Hawaii National Park, is on the summit of the dormant volcano Haleakala which rises to a height of 10,025 feet and comprises the east part of the island of Maui, Territory of Hawaii (Pl. 1). The mountain is a broad dome or shield-shaped mass about 34 miles from east to west and 25.5 miles from north to south (Fig. 3). Evidence is presented to show that it is two valley heads and not a crater, caldera, or eroded caldera. It will be referred to here as the summit depression. The depression is 7 miles long from east to west and 2 miles wide from north to south. At the east end it is notched by Kaupo Valley which drains to the south and at the west end by Keanae Valley which drains to the north. The gap at the head of Keanae Valley is known as Koolau Gap. The walls of the summit depression between the gaps at the heads of these valleys vary in height from 560 to 1800 feet above the floor. The floor is formed of cinder and spatter cones from a few feet to 650 feet in height and of fresh-looking black and brown basalt flows, chiefly aa. Extending from the summit are three rift zones marked by cone and crater chains (Fig. 3). They trend N. 80° E., N. 30° W., and S. 65° W., respectively. A prominent cone and crater chain built over the southwest and east rifts crosses the depression floor (Fig. 3; Pl. 1).

## PREVIOUS INVESTIGATIONS

Pickering and Drayton of the Wilkes Expedition explored the summit depression of Haleakala, and Drayton's sketch of it is apparently the first published detailed map (Dana, 1849, p. 228). Dana surmised, without visiting the depression, that Kaupo and Keanae valleys probably marked the sites of great ruptures attending the last summit eruption which caused the great eastern segment of the volcano to be separated from the western by the width of these valleys. His rent hypothesis has been accepted by many later geologists. As a result of a visit in 1887, Dana (1890, p. 278) set forth another view—namely, that these valleys might be grabens.

Daly (1914, p. 153) was inclined at first to accept the rent hypothesis but subsequently he expressed the idea that the depression might be due to both renting and slumping (Daly, 1933, p. 171). Dutton (1884, p. 105) states that the depression is strictly homologous in origin to Kilauea

caldera. Daly (1933, p. 171) disagrees with this theory of origin and states that general circumferential faulting, which is topographically so clear at Kilauea, is not evident on the walls of the vast depression of Haleakala.

Cross (1915, p. 92-93) cogently remarks that:

> "What is commonly called the crater of Haleakala appears to me to be in some part at least, a result of erosion. East of the borders of the flow through the Kaupo gap are canyons which do not head so far back as the crater but whose beds are deeper than the surface of the flow. These canyons and many others shown by the map indicate the character of the gorge which once penetrated through the Kaupo gap into the summit crater. How much of the crater is erosional remains to be determined."

Sidney Powers (1917, p. 512) states:

> "Cross argues that the Koolau and Kaupo gaps were formed by valley erosion rather than by block-faulting. If so, why does the Keanae Valley (Koolau Gap) grow smaller downstream when it heads in a trench so dry that no vegetation grows? The Kipahulu rifts are pronounced, according to authentic reports, near the Haleakala rent, and certainly the walls of the Kaupo Gap do not represent products of erosion."

Hinds (1931, p. 174) and Powers (1917, p. 512) favor Dutton's caldera theory for the formation of the depression and Dana's graben theory for the formation of the gaps, but both state that further study is needed to determine the nature of the displacements.

W. O. Clark, the first geologist to make detailed studies of East Maui, wrote in an unpublished manuscript, seen by the writer in 1924, that he believes Keanae Valley is due to erosion and that it tapped a large subsidence caldera.

### OBJECTION TO THE RENT HYPOTHESIS

An objection to the rent hypothesis is that the east segment of the dome, if pushed westward an amount equal to the width of the gaps, would not fit the western portion. The depression walls at the head of Keanae Valley are 3 miles apart, and 5 miles down the valley they are only three quarters of a mile apart; hence it is obvious that pushing the two segments together so that the walls fit would cause one side of Keanae Valley to overlap the other side by about 2 miles. Subsequent erosion in the head of the valley might account for part of the misfit but not for all of it. Considering the fact that Dana based this hypothesis upon inspection of a rather crude map, which was certainly suggestive of the rent hypothesis, it is not surprising that he set forth a new hypothesis when he later saw the depression.

### OBJECTIONS TO THE GRABEN HYPOTHESIS

If Kaupo and Keanae valleys are grabens caused by the subsidence of two elongated blocks, then the great amphitheater-headed Kipahulu

and Waihoi valleys adjacent to Kaupo Valley remain to be explained (Figs. 2, 3). Powers (1917, p. 5) believed Kipahulu Valley was a graben and based this statement on "authentic reports." However, neither he nor the other geologists writing about the depression indicate that they ever saw Kipahulu and Waihoi valleys.

Kipahulu and Waihoi valleys, except for being partly filled with late lavas, are typical amphitheater-headed valleys characteristic of Hawaii. They are narrow near the sea and wide at their heads (Fig. 3). The walls at their heads rise 2000 feet above their lava-plastered floors and before the late lava poured into them they were more than 1000 feet deeper. A narrow divide at the eastern edge of the crater separates the head of Kaupo Valley from Kipahulu Valley (Pl. 1). A similar divide separates Kipahulu Valley from Waihoi Valley. Because sufficient time elapsed for the erosion of the two great valleys of Kipahulu and Waihoi, the writer agrees with Cross that Kaupo and Keanae valleys are more readily explained by stream erosion than by faulting.

## OBJECTIONS TO THE CALDERA HYPOTHESIS

Daly has already pointed out the absence of circumferential faulting as an objection to Dutton's statement that the crater is a collapse caldera. The absence of such faulting is, however, not conclusive proof, as some calderas of collapse are not bordered by arcuate faults, and erosion might remove evidence of such faults if they had been present. The walls of the summit depression consist of numerous cinder cones, plugs, dikes, and fire-fountain deposits interstratified with narrow, irregular flows, chiefly aa, dipping 10 to 35 degrees away from a projection of the present crest line of the mountain. The rocks are mostly basaltic andesites and andesites similar to those found along the present rift zones and are vastly different from the even-bedded basalts, chiefly pahoehoe, found in the basement of Haleakala or in the walls of Kilauea and Mauna Loa calderas on Hawaii. Although it probably cannot be proved that calderas did not persist on the summits of the eroded Hawaiian volcanoes capped with thick deposits of cinder cones and flows of andesitic type, it can be stated with assurance that the slightly eroded cone-studded Hualalai and Mauna Kea volcanoes, both of this type, do not have calderas and that the basaltic volcanoes, Kilauea and Mauna Loa, have calderas. It appears that when Hawaiian volcanoes stop erupting typical highly fluid primitive olivine basalts and start producing massive and somewhat differentiated lavas, the cinder cones and lavas concurrent with the latter phase may be sufficiently numerous and bulky to fill

and obliterate calderas. Certainly the lava beds making up the walls of the summit depression of Haleakala stand in strong contrast to those in the basement of Haleakala Volcano and to those in Kilauea and Mauna Loa. The gentle dips of 2 to 6 degrees in the lava beds sur-

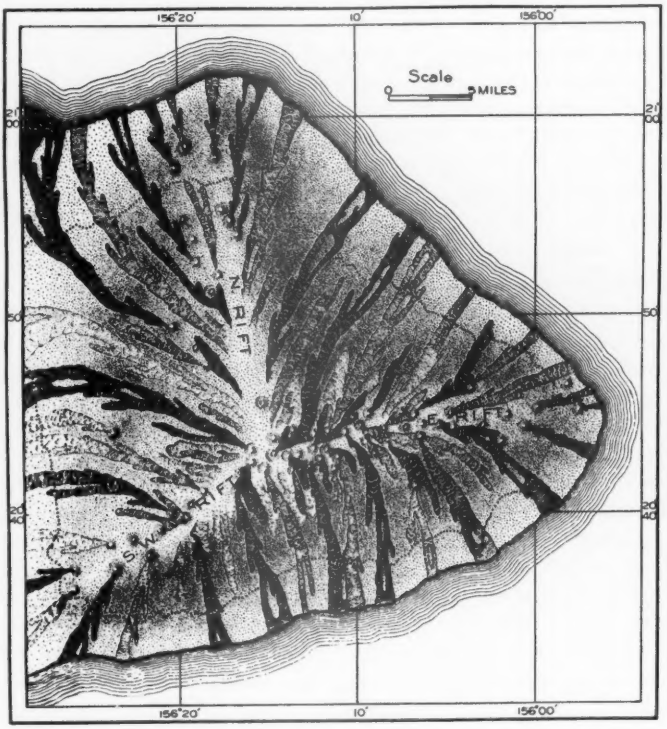


FIGURE 1.—*Haleakala Volcano prior to the erosion of the great valleys and prior to the great submergence*

The dotted line shows the present shore line.

rounding Hawaiian calderas indicate accumulation on a sunken summit, and the even flatter-lying, ponded lavas of the caldera floors give unmistakable evidence of accumulation within a great depression. Such flat dips are notably lacking in the walls of Haleakala. Instead, the dips are 10° to 35° and are similar to those on adjacent rifts. The probable appearance of Haleakala Volcano prior to the erosion of the great valleys is shown in Figure 1.

## PRESENT INVESTIGATIONS

The writer photographed a dike swarm in the depression floor just east of Koolau Gap during his first visit in 1924. The exposure of this dike swarm on a low ridge suggested that appreciable erosion must have occurred prior to the partial mantling of these dikes by recent fire-fountain deposits and lava. From 1932 to 1938 the writer resided on Maui and made numerous trips to the summit of Haleakala while systematically mapping the geology and ground-water resources of Maui for the United States Geological Survey in financial co-operation with the Territory of Hawaii. The mapping of the floor and gaps was completed in September 1935, and 3 months were spent in 1940 studying the walls of the depression in detail and the adjacent Waihoi and Kipahulu valleys.

## EROSIONAL HYPOTHESIS

The hypothesis is presented that the so-called crater is a product of stream erosion resulting from the coalescence of Keanae and Kaupo amphitheaters, each having receded headward from opposite sides of the mountain (Fig. 2). Small craters, similar to those on the southwest rift zone, may have been present, but a large caldera resembling the present depression did not exist at the time these canyons reached the crest. The large size of the depression is chiefly the result of the valley heads being offset and not uniting in a straight line across the top of the mountain. The hypothesis is based on the following observations: (1) A depression exactly the shape of the so-called crater is formed if one reconstructs amphitheaters at the heads of Kaupo and Keanae valleys analogous with those of the adjacent Kipahulu, Waihoi, and Manawainui valleys (Fig. 2); (2) the walls of the depression have suffered appreciable erosion as shown by the exposure of numerous dikes, partly dissected cones, and small canyons; (3) a ridge which appears to be a remnant of the divide that formerly separated the two valley heads is present in the depression floor extending south from Hanakauhi Peak. The ridge is cut by closely spaced dikes and plugs which now stand in relief; (4) a stream once reaching the rift zone would have expanded its head rapidly because of the loosely knit structure resulting from the presence of so many weak cinder cones; (5) a dike complex that must have yielded perennial spring water which would have accelerated erosion crops out at the heads of both valleys.

Landsliding has flattened the walls of the former heads of Keanae and Kaupo valleys now forming the so-called crater walls as shown by the voluminous talus fans that interfinger with the late flows in the crater.

A coarse breccia rises 355 feet above the coast at the mouth of Kaupo Valley. Its extension below sea level is unknown. It is separated at the top by an erosional unconformity from the plastering lava flows in

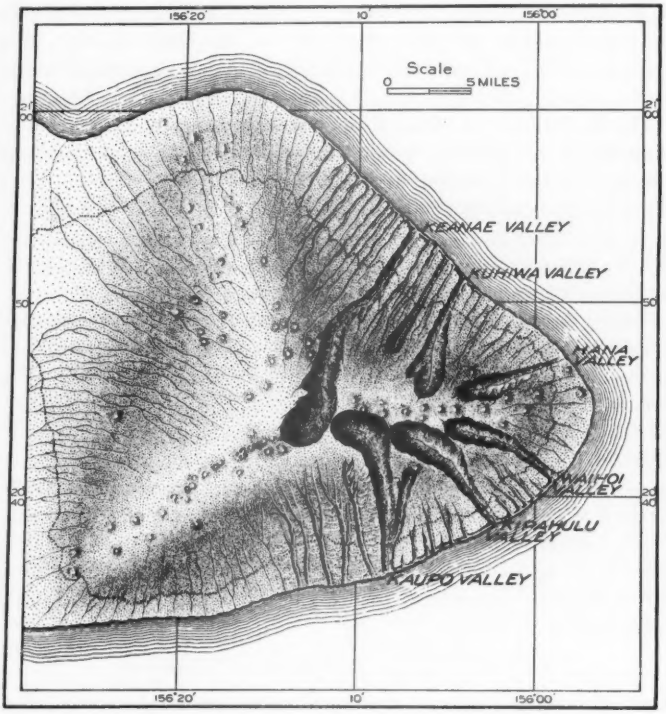


FIGURE 2.—*Haleakala Volcano after the erosion of the great valleys and prior to the great submergence*

The form of Kuhiwa and Hana valleys is largely conjectural, as they are now deeply buried by later flows. The dotted line shows the present shore line. Manawainui Valley is the small, deep, unnamed tributary of Kaupo Valley.

Kaupo Valley and in places is covered with conglomerates. It appears to be a mud flow that moved down Kaupo Valley prior to the late eruptions as a result of the great talus deposits bordering the valley walls being set into motion following heavy rains. Boulders of similar breccia were found in the alluvium at the mouth of Keanae Valley, which may indicate the existence of similar mud-flow deposits of the same age in that valley also.

The walls of the depression have been carefully searched for faults, and evidence was found of only one. The fault escarpment lies stratigraphically under about 1000 feet of lavas at the base of the south wall. It cannot have caused the summit depression because the downthrown block is south of the rim. The fault cliff truncates porphyritic pahoehoe basalt that strikes N. 75° W. and dips 8° S. The cliff trends E.-W. and dips 83° S. and has along its surface 2 feet of talus composed of vesicular pahoehoe mixed with red ashy soil. The rock is weathered beneath the soil. On the soil rest aa lavas of the andesitic group striking E.-W. and dipping 35° S. A lens of conglomerate and hillwash was found interstratified with the overlying andesitic lavas. The buried cliff is interpreted as a south-facing fault escarpment more than 500 feet high, over which about 1000 feet of lavas from the summit ridge cascaded long before the great valleys were cut. The conglomerate was deposited by short-lived streams rising on this ridge.

The fault caused a shortening of the south slope of the mountain but is buried too deeply to have played a part in the development of the summit depression. However the facts that about 1000 feet of lava flowed from the north over it and that streams laid down gravel are proof that a caldera the size of the present summit depression did not exist during the accumulation of the andesitic lavas.

Lenses of gravel were found between flows in the north walls of the depression, indicating that streams flowed from a ridge that existed where the present summit depression lies. The south rim of the present depression is notched about midway by a valley that formerly drained to the south from this projected ridge and was subsequently beheaded by Keanae Valley.

## OBJECTIONS TO EROSIONAL HYPOTHESIS

The following objections to the erosion theory might be raised: (1) The east rims of Keanae and Kaupo valleys are lower than the west rims; (2) the size of Keanae Valley is abnormally large in comparison with the valleys on both sides of it; (3) the valleys west of Kaupo Valley are small; (4) the low precipitation on the summit, especially at the head of Keanae Valley, is inadequate to cut such a large canyon; (5) the course of Keanae Valley changes from north to northeast at an altitude of 2500 feet; (6) the width of Keanae and Kaupo valleys increases headward; (7) fresh-looking flows form part of the rim of the depression. Answers to these objections are offered here:

(1) The east rims of Keanae and Kaupo valleys are lower than the west rims, due to the streams eating headward, not into a symmetrical cone but into a ridge that slopes down toward the east along its crest.

This is typical of the other large valleys. For example, the highest point on the west side of the head of Kipahulu Valley is 8101 feet, and the highest point on the east side is 5550 feet. Respective altitudes at the head of Waihoi Valley are 5550 feet and 4400 feet.

(2) The size of Keanae Valley is abnormal in comparison with the valleys on both sides of it. W. O. Clark, who since 1929 has been supervising geologist of water developments for the East Maui Irrigation Co. in the vicinity of Keanae Valley, has demonstrated by numerous tunnels that many lava-filled valleys lie east of Keanae Valley, and that the lava partly filling Keanae Valley occupies numerous secondary valleys cut during the intervals between lava flows. At the mouth of Keanae Valley, older alluvium forms a terrace about 100 feet above sea level, and this alluvium underlies all the exposed valley-filling lavas and in turn fills a wide valley cut considerably below sea level. This big valley was evidently cut to some low stand of the sea during the Pleistocene, as were the other major valleys in the Hawaiian Islands, and later filled with alluvium during stands of the sea higher than the present (Stearns, 1935). Subsequent dissection of the alluvium flooring the valley has been interrupted by lava flows from the summit depression. The details of the volcanic accidents to these streams will be described in the forthcoming Bulletin 7 of the Hawaii Division of Hydrography.

The volume of late plastering lavas from Haleakala has been so great in the area east of Keanae Valley that most of the topographic evidence of former valleys has been obliterated except at the head of Kuhiwa Stream above Nahiku and just west of Kawaipapa Gulch near Hana. A similar, although much less voluminous, veneer of lavas covers the country west of Keanae Valley.

(3) The small size of the valleys west of Kaupo Valley in comparison with the large size of the valleys to the east is due to the rapid decrease in rainfall from east to west owing to the trade winds being shut off from that area. The valleys in that area have also been heavily plastered by lava flows, like those between Hana and Keanae. Manawainui Gulch, a tributary of Kaupo from the east that was not plastered by later lavas, is a steep-walled amphitheater only 3200 feet wide and yet 3200 feet deep at its head. This spectacular amphitheater indicates the intensity of erosion in this area where not interrupted or covered by later lavas (Fig. 3).

(4) The low precipitation at the head of Keanae Valley has been pointed out by Powers (1917, p. 512) as a serious objection to the erosional hypothesis. Evidence was found on Oahu that during the glacial periods the rainfall belt in the Hawaiian Islands moved lower and that

during interglacial periods it moved higher than at present (Stearns, 1935, p. 1953). Thus, during interglacial periods warmer than the present, the rainfall should have been considerably greater on the top of the mountain than it is now. Precipitation on the summit now averages 40 inches annually.

(5) The change in the course of Keanae Valley from north to northeast at an altitude of 2500 feet might be cited as indicating some influence of drainage from a summit crater on a normal northeast course for Keanae Valley. This change in course appears entirely due to the topographic influence of the high land along the north rift zone of the mountain prior to dissection. Honomanu Stream just west of Keanae Valley changes its course exactly at the same place on the mountain slope and parallels Keanae Valley throughout its course, even though it does not reach back to the summit depression.

(6) The great increase in the width of Keanae and Kaupo valleys headward has been cited by Powers (1917, p. 512) as an objection to the erosional origin of these valleys. The expansion of major valleys headward into great amphitheaters is characteristic of Hawaiian valleys (Stearns and Vaksvik, 1935, p. 24-26). That both Keanae and Kaupo valleys tapped dike complexes is indicated by the presence of more than 70 dikes exposed in the walls at the head of Keanae Valley and 60 at the head of Kaupo Valley. It is of course not known how much water was present in such dike structures at the time the valleys were cut. Perennial ground water supplied by similar complexes has been shown to be very effective in increasing the erosion of such large valleys (Stearns and Vaksvik, 1935, p. 25), and it is probable that both valleys contained perennial springs in their headwalls. Waihoi and Kipahulu valleys likewise penetrated the dike complex and were similarly benefited. Further, all these valleys encountered weaker and more permeable rocks as they approached the loosely knit structure of cinder cones and highly vesicular and cavernous irregular lava beds of the rift zone complex now well exposed at the heads of the valleys. The major valleys all cut down to primitive olivine basalts which yielded rapidly to stream erosion. The lava flows erupted on the Haleakala rifts have commonly flowed down the rifts for some distance before plunging seaward down the main slope of the mountain. This is shown by lava beds in the south wall of the summit depression between the summit and the saddle midway to Kaupo Gap. In this stretch of cliff the lava beds have a definite east dip from the summit toward the saddle as well as a dip to the south. The east component of the dip must have caused more surface drainage to enter the west side of the heads of the valleys and thereby caused them to expand

in that direction. This effect is well shown by the westward lobes on Kipahulu and Waihoi valleys (Fig. 3).

Supplementing these causes for the wide heads of the valleys is the fact that the side-wall drainage now flows seaward along the contact

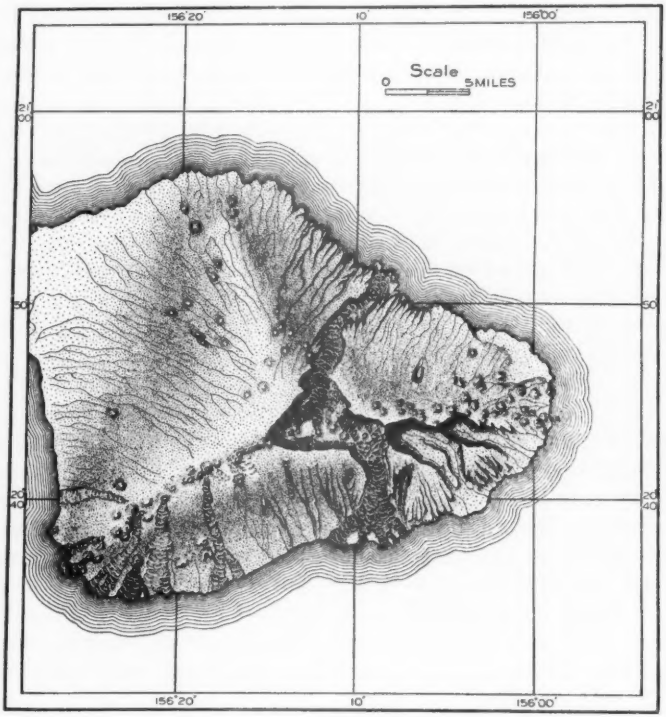


FIGURE 3.—*Present-day Haleakala Volcano after numerous secondary eruptions have filled or partly filled the great valleys and after the great submergence*

of the walls and the late lavas flooring these valleys. Each valley is therefore occupied now by two streams instead of one. Thus, deepening of the original valleys ceased soon after the first flows of lava, and subsequently only lateral under cutting of the side walls has occurred. Almost every lava flow crowds these streams farther toward the walls. Thus the original depth of the valleys cannot be determined by simply projecting down the profiles of the valley walls to the point of intersection.

Keanae Valley would naturally have been the largest valley on the windward northeast slope of Haleakala, because it drained the largest area—the re-entrant between the northwest and east rift zones. Westward from Keanae Valley the length and number of tributaries of each main stream decrease rapidly (Fig. 3). The amount of plastering lavas erupted from the north rift has been small compared with the other rifts; hence this rapid decrease in drainage area is probably a more effective cause of the small size of valleys west of Keanae Valley than the mere plastering of stream valleys.

(7) The fresh-looking lava flows forming part of the rim of the summit depression might be cited as proof that lava lakes recently existing in it overflowed the rim before the floor collapsed. The fresh lavas on the rim can be traced in several places to veneers which lie unconformably on the depression wall, conclusive evidence that they were poured from fissures after the depression had reached its present form. A fine example of a fissure which poured out lava simultaneously on the floor, on the wall, and on the rim occurs at Puu Hanakauhi, the peak on the east side of Keanae Valley.

## SUMMARY

No stratigraphic or structural evidence was found to support the hypotheses that Haleakala Crater is a true caldera, that it was formed by renting, or that Keanae and Kaupo valleys tapping this depression are grabens. Instead, detailed mapping and the examination of water tunnels show that Haleakala dome has been eroded by a number of great valleys. The hypothesis is presented that the so-called "crater" of Haleakala is chiefly, if not entirely, the result of the coalescence of the amphitheater heads of Kaupo and Keanae valleys and that renewed volcanic activity has partly masked their former divide and partly filled these valleys with lava flows.

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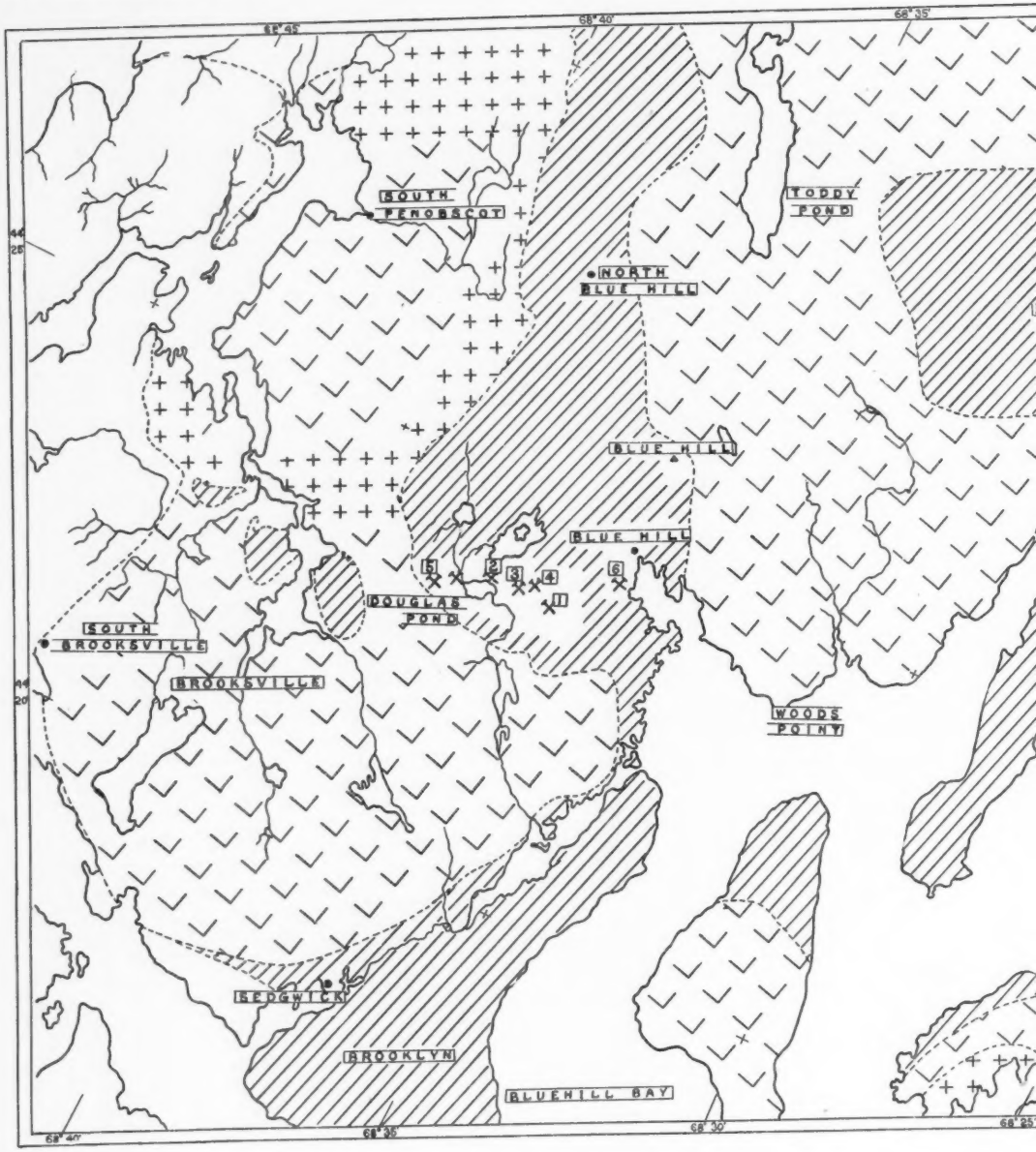
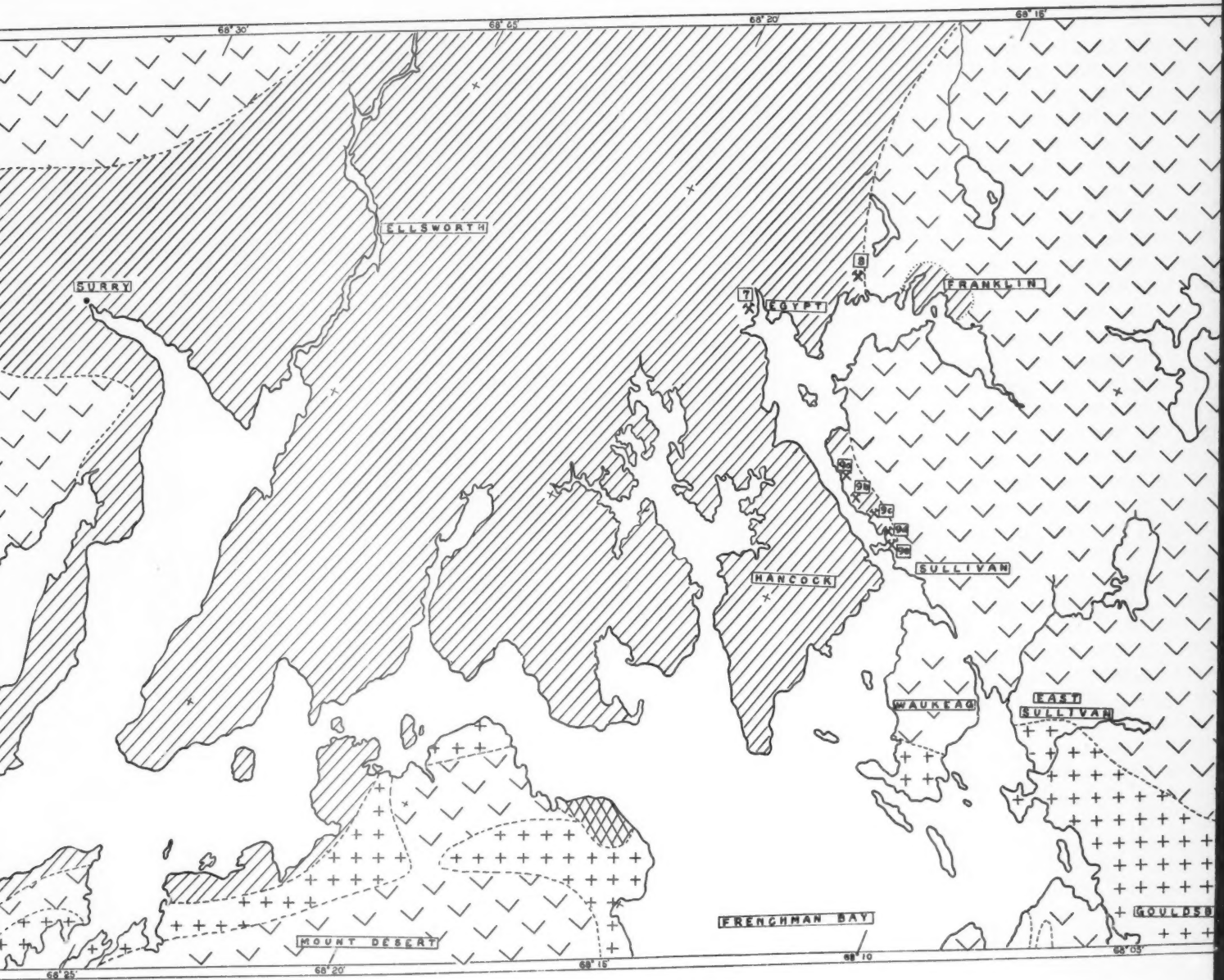
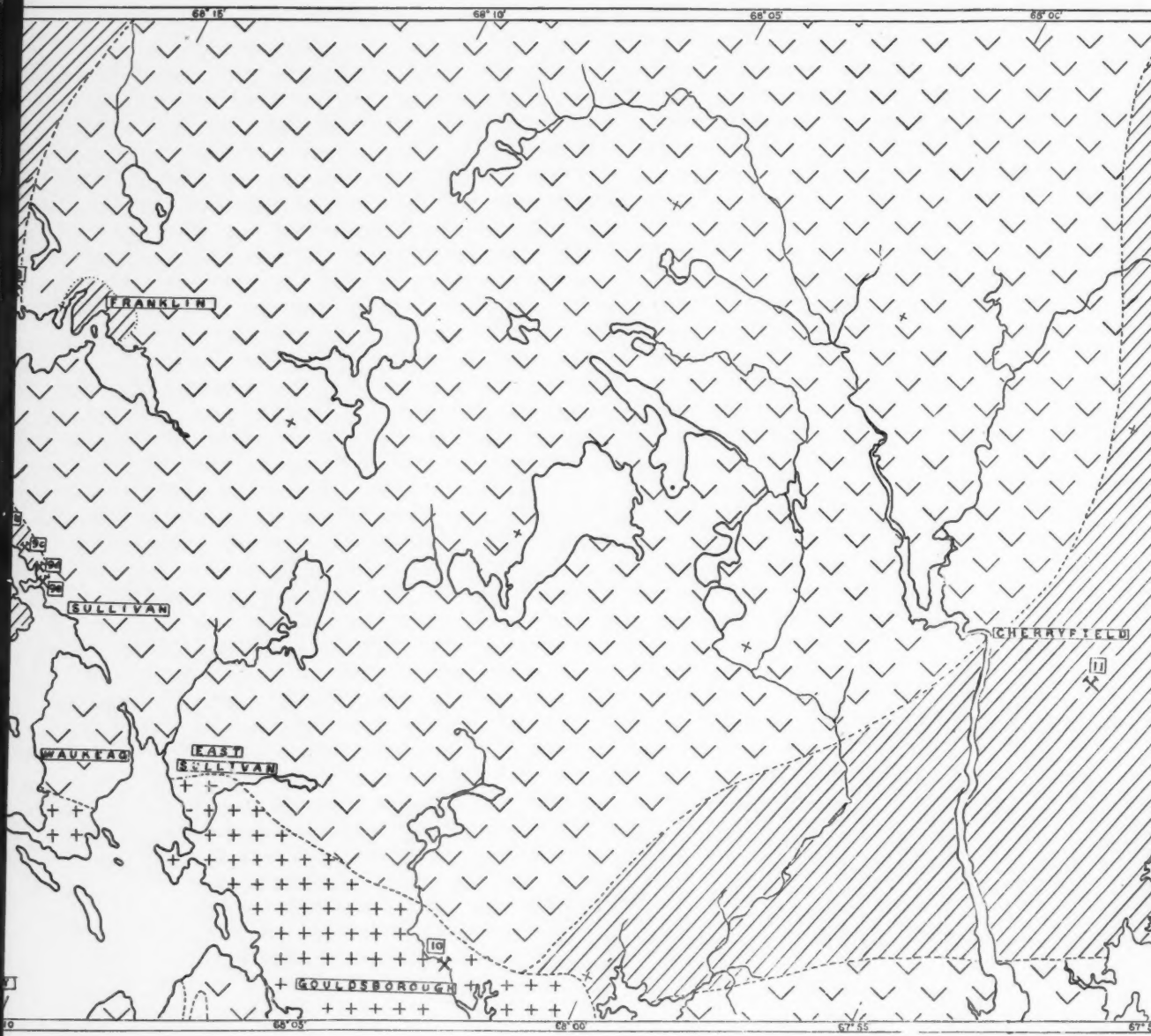
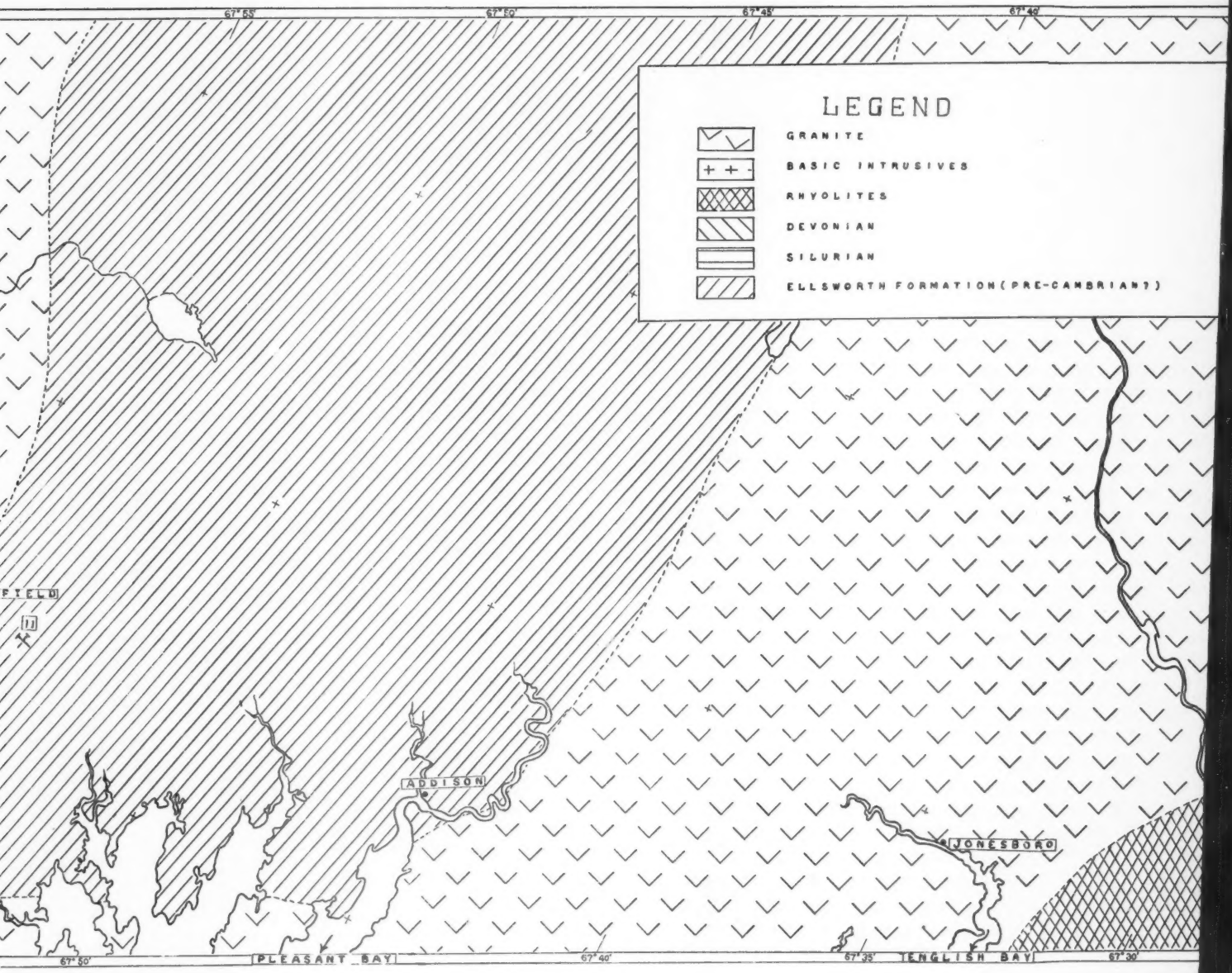

68°45'
68°40'
68°35'
SOUTH PENOBSCOT
NORTH BLUE HILL
TODDY POND
BLUE HILL
BLUE HILL
DOUGLAS POND
SOUTH BROOKSVILLE
BROOKSVILLE
WOODS POINT
REDGWICK
BROOKLYN
BLUEHILL BAY
68°40'
68°35'
68°30'


68° 30'
68° 25'
68° 20'
68° 15'
68° 10'
68° 05'
ELLSWORTH
SURRY
EGYPT
FRANKLIN
HANCOCK
SULLIVAN
WAUKEAG
EAST SULLIVAN
GOULDSB
FRENCHMAN BAY
MOUNT DESERT
7
8
9a
9b
9c
9d
9e



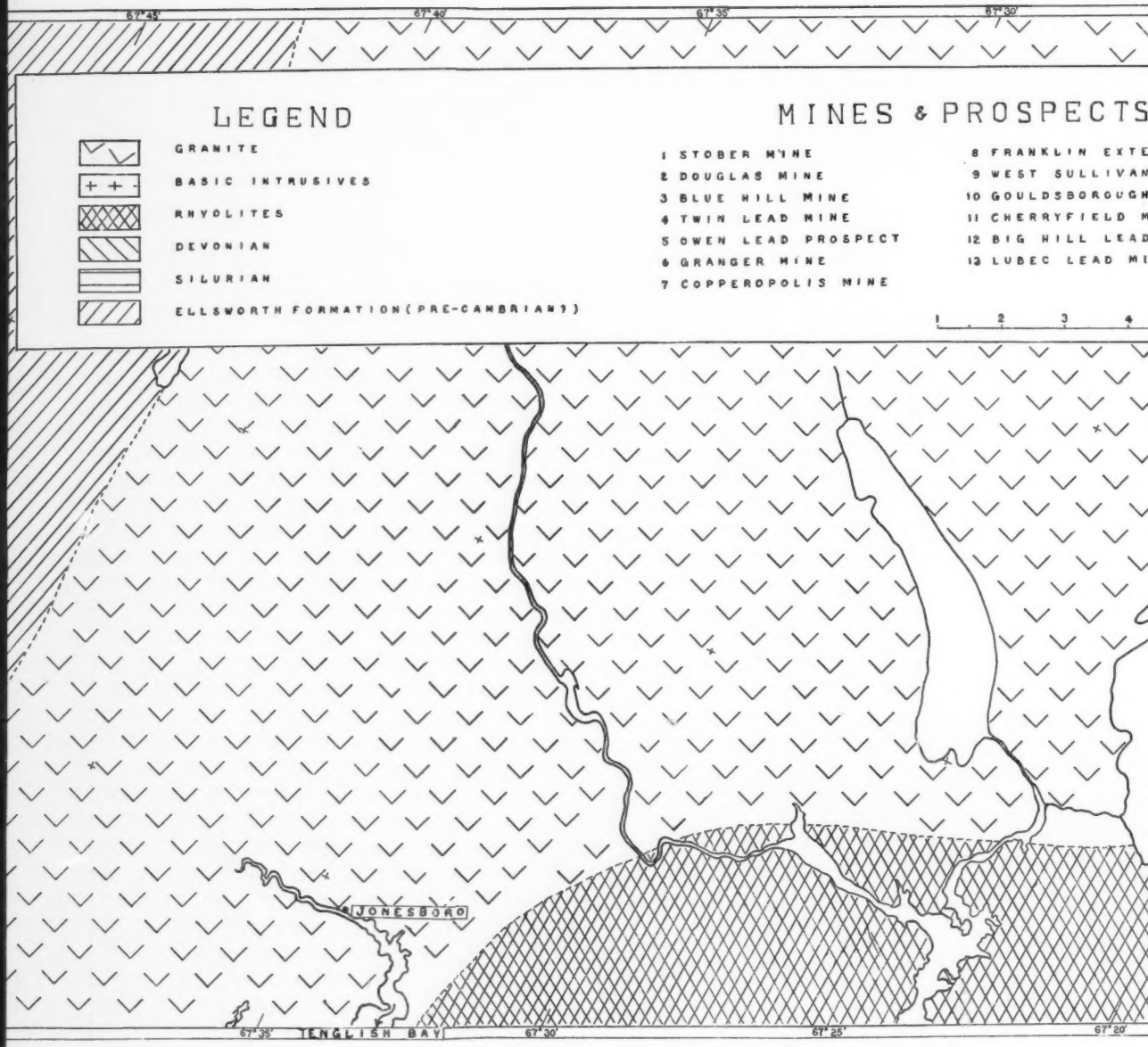

GENERALIZED GEOLOGIC MAP OF

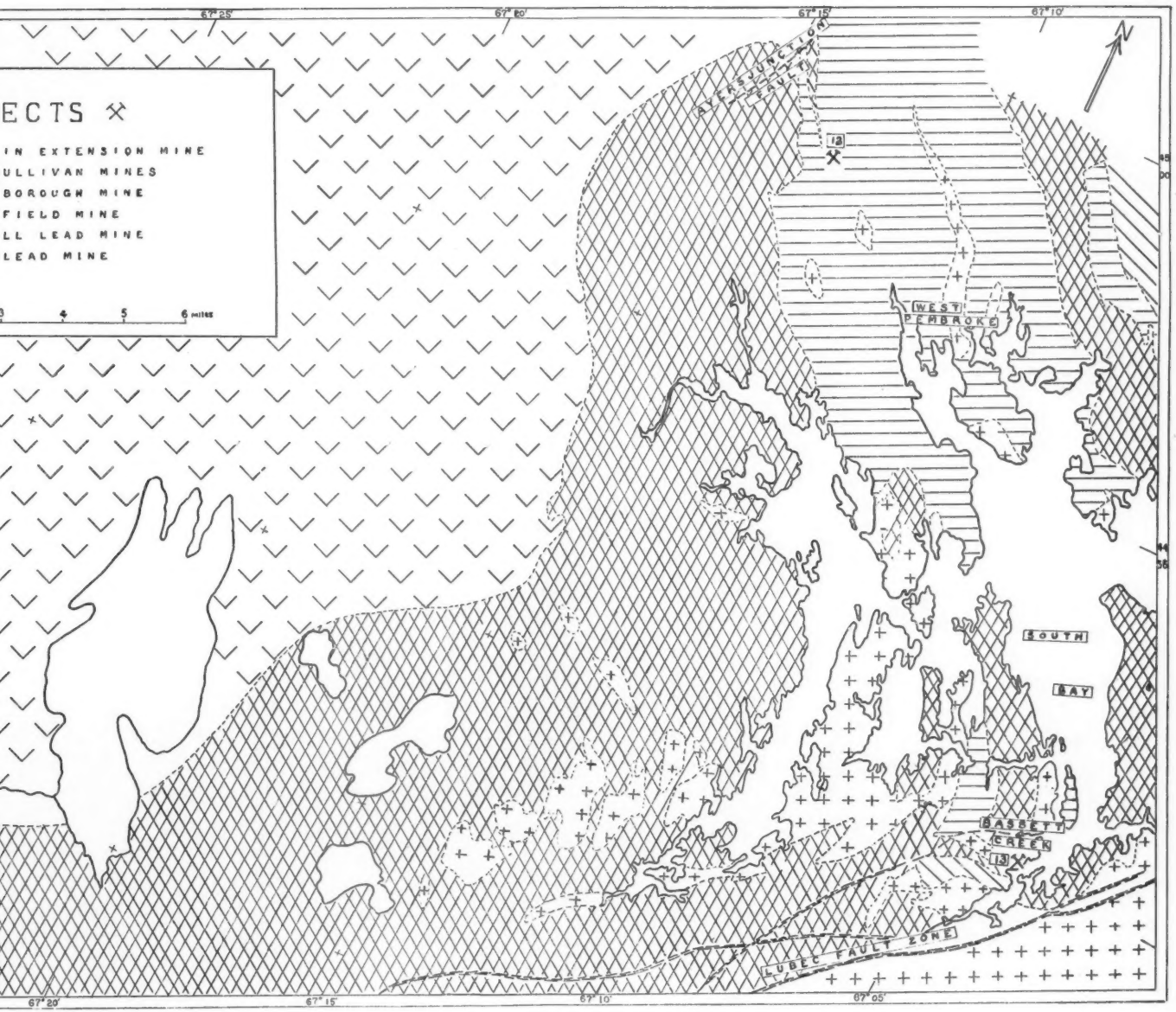
Based largely on works of G. O. Smith, E. S. Bastin, A. Keith,




MAP OF SOUTHEASTERN MAINE

n, A. Keith, C. W. Browns, H. S.Williams, with modifications.


67° 45'
67° 40'
67° 35'
67° 30'
LEGEND
GRANITE
BASIC INTRUSIVES
RHYOLITES
DEVONIAN
SILURIAN
ELLSWORTH FORMATION (PRE-CAMBRIAN?)
MINES & PROSPECTS
1 STOBER MINE
2 DOUGLAS MINE
3 BLUE HILL MINE
4 TWIN LEAD MINE
5 OWEN LEAD PROSPECT
6 GRANGER MINE
7 COPPEROPOLIS MINE
8 FRANKLIN EXTE
9 WEST SULLIVAN
10 GOULDSBOROUGH
11 CHERRYFIELD M
12 BIG HILL LEAD
13 LUBEC LEAD MI
1 2 3 4
JONESBORO
67° 35'
TENGLISH BAY
67° 30'
67° 25'
67° 20'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ECTS
IN EXTENSION MINE
ULLIVAN MINES
BOROUGH MINE
FIELD MINE
LL LEAD MINE
LEAD MINE
4
5
6 miles
WEST PEMBROKE
SOUTH BAY
BASSETT CREEK
LUBEC FAULT ZONE
13
13

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 15-52, 5 PLS., 1 FIG. JANUARY 1, 1942

# GENESIS OF SOME ORE DEPOSITS OF SOUTHEASTERN MAINE

BY CHING-YUAN LI

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## ABSTRACT

The paper intends to show the genetic relation between the ore deposits and the granite intrusive of southeastern Maine. The deposits are ores of copper, zinc, lead, and a little silver, occurring mostly in the so-called Ellsworth schist, and some in the diorite and Silurian rocks. Nearly all of them occur from ¼ to 2 miles from the contact with the granite. A rude zonal arrangement of the deposits around the granite is noticed. Their genetic relation is further confirmed by the paragenesis of ore minerals. The sequence is apparently as follows: magnetite, arsenopyrite, pyrite, pyrrhotite, sphalerite, chalcopyrite, and galena. This sequence holds true in general in spite of the fact that the mines are scattered over more than 100 miles. The influence of the granite intrusion upon the surrounding rocks is studied in some detail. Particular emphasis is put on the petrographic description of the thermal metamorphism imposed on the schist. A short description is given for each mine and prospect.

## INTRODUCTION

### STATEMENT OF THE PROBLEM

Nearly 70 years ago, during a mining "boom" along the southeastern coast of Maine, mining companies organized to exploit the various copper, lead, and zinc deposits of the region. Operations were soon suspended after the ores proved to be of low grade. Since then, few geologists have paid much attention to these deposits. More than 30 years ago, Emmons (1910) published a paper on this region, in which he classified the ore deposits into three groups: (1) those formed before regional metamor-

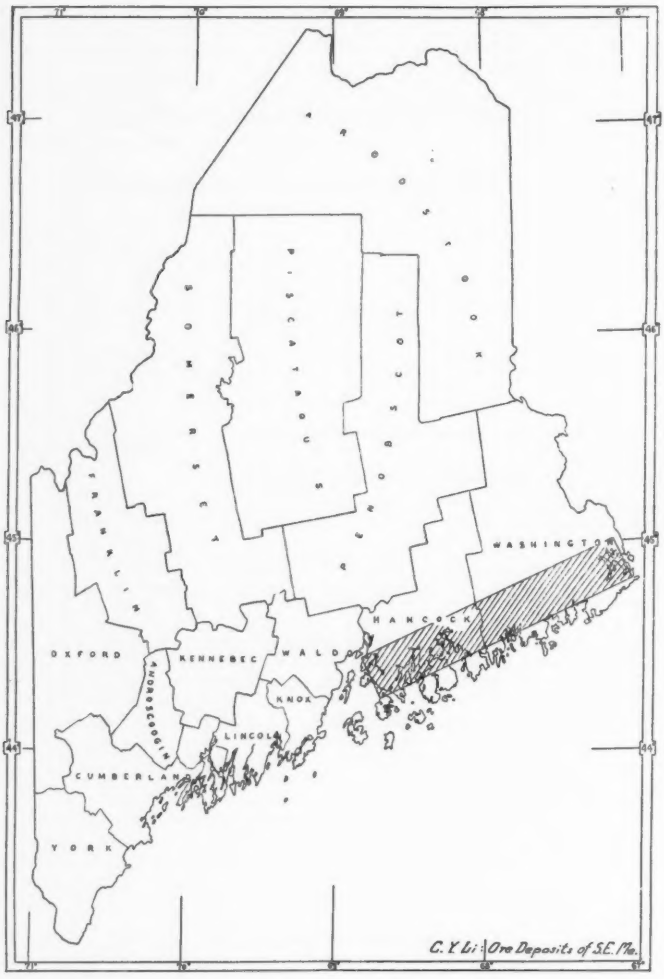



FIGURE 1.—*Location map*

phism; (2) those associated with the granite intrusion; and (3) those associated with diabase and traps.

The present writer believes that all the deposits are genetically related to the granite intrusion; that there is a rude zonal arrangement of the deposits dependent upon their relative distance from the granite; and that

those deposits considered by Emmons as of pre-metamorphic origin were actually formed by fluids expelled from the granite.

### BASIS OF REPORT

This report is based upon a year's continuous study which involved field, petrographic, and mineragraphic investigations. Field work was done in the summer of 1940, when the writer visited the various mines and prospects. Laboratory work was carried on at Columbia University.

### LITERATURE

Important scientific works dealing with the geology and ore deposits of the area have been few. Eastport Quadrangle has been mapped by Bastin (1914). Blue Hill district is included in the Penobscot Bay Quadrangle (Smith and Bastin, 1907). A preliminary geologic map of the State of Maine was compiled by Keith (1933a). Emmons' paper (1910) is the only one which has dealt with the ore deposits of the whole area.

The pyritic copper deposits of the Blue Hill district have received some attention. D. G. B. Thompson (1923) first questioned Emmons' explanation of the deposits of that part of the area. This study aroused the interest of Lindgren (1925) who published a short description of the occurrence. Gillson (1929) has described in some detail the contact metamorphism imposed on the schist by later granite at Blue Hill. Newhouse (1930), in connection with his work on the sulphide ores from the Blue Hill district, studied them on polished surfaces and concluded that they are later than the regional metamorphism.

### ACKNOWLEDGMENTS

The problem was suggested by the late Dr. Philip K[illegible]eger who spent a few days with the writer in the field.

The writer is much indebted to Professor William Agar, under whose direction the work was completed, and to Professors S. James Shand and Walter H. Bucher for their critical reading of the manuscript. Dr. Eugene Cameron checked the mineragraphic studies. Mr. Eric Lloyd Evans was very helpful in the instrumental work.

Gratitude is due Professor Freeman Burr, State Geologist of Maine, who accompanied the writer to Augusta to make arrangements with the State officials in order to make geologic surveying work possible for a foreigner.

## GEOLOGY AND PETROGRAPHY

### GENERAL STATEMENT

Granite, with minor quantities of diorite and diabase, occupies approximately two thirds of the area under consideration (Pl. 1). It intrudes

both the Ellsworth formation, which is next in abundance, and a series of Silurian rocks. Since pebbles of this granite are present in the conglomeratic facies of the Perry formation in the Perry Basin, which has been determined to be Upper Devonian from its plant remains (Smith and White, 1905), the granite must have been intruded sometime during late Silurian to middle Devonian time.

The Ellsworth, composed chiefly of schist, is the oldest formation in this area. It is definitely older than the Silurian and has been commonly regarded as pre-Cambrian due to its intense metamorphism (Keith, 1933a). Its total thickness is unknown.

Silurian shale grading in places into well-stratified tuff is interbedded with volcanics, principally rhyolite tuffs and flows. The total thickness is estimated at about 14,000 feet (Bastin, 1914).

The major structural features of the area arrange themselves in a NE.-SW. direction parallel to the regional structural pattern of Maine. Separated by other rocks into several disconnected areas, the granite intrusion, if viewed as a whole, forms a NE.-SW. belt. Local variation of the strike of the schistosity of the Ellsworth formation is very marked, but the general strike is also about NE.-SW. The Silurian rocks form a gentle northeastward-pitching anticline in the east bordered by a NE.-SW. striking fault on either side.

The granite and the Ellsworth formation are discussed more fully than other rocks, since the former is considered to be genetically related to the ore deposits, and the later is the country rock of most of the mines.

### GRANITE

*Geographic distribution.*—Granite, the most widespread rock of the region, outcrops parallel to the southwest-northeast-trending coast from Penobscot Bay up to Jonesboro in English Bay. Within this belt are included several large islands and the mountains of Mount Desert Island which owe their topographic prominence to the massive granite. This coastal belt turns northward at Pleasant Bay near Addison, then curves gradually northeastward across St. Croix River to New Brunswick, Canada. Two other areas of granite, trending roughly north-south, extend northward from the northern end of both Frenchman Bay and Blue Hill Bay.

*Description.*—The great mass of granite is gray to pinkish gray, medium-grained, with orthoclase, microcline, oligoclase, quartz, and biotite as its chief constituents. The remarkable uniformity of the mineral composition over a considerable area is seen from the published descriptions of specimens from different quarries in and near the region.

*Contacts.*—The marginal phase of the granite shows marked departure from the normal texture. It is generally porphyritic, and where the granite has injected sedimentary rocks the resulting complex is gneissoid, as the exposures in Sullivan and near the Franklin Extension Mine show.

Hand specimens from contact outcrops often show fairly well on a small scale the different ways in which granite has intruded the surrounding rocks. Certain granite borders are very sharp; others are indefinite. In the vicinity of Blue Hill and Brooksville, the contact is sharp, pure granite being followed within a foot by schists entirely free from granite. On the shore, half a mile west of Woods Point near Blue Hill village, there is considerable intrusion of granite parallel to the foliation of the schists. Due east of North Blue Hill, half a mile west of Toddy Pond, dikes of granite parallel the foliation of the schist and also break across it and have caught up a number of angular fragments of schist.

Around South Penobscot, the granite area is completely surrounded by a border zone from ¼ to 1½ miles wide in which the rocks are largely diorite, with some gabbro. Throughout this border zone some granite occurs even at the outer border.

*Petrography.*—Granite from South Brooksville, which is thought to be representative of the bulk of the granite of this region, was chosen for microscopic study. It is a medium-grained, light pinkish-gray biotite granite, with an allotriomorphic interlocking texture. Three chief constituents—feldspar, quartz, and biotite—compose about 60, 30 and 6 per cent, respectively, of the rock. The richness in quartz and relative poverty in biotite is worth noting. The feldspar and quartz crystals may be as large as 1.6 mm. across. Quartz in small grains occupies the interstices. It is usually very clear. Microcline and orthoclase are the dominant feldspars with subordinate amounts of oligoclase. Microcline is the most abundant and is perthitic. These feldspars are partially sericitized and kaolinized. Biotite is dirty green to dark brown, highly pleochroic, with zircon inclusions surrounded by dark haloes. The ratio between the width and length of the biotite flakes averages about 1:1.9. The biotite is occasionally altered to chlorite and magnetite. Apatite is another minor constituent. Normal sequence of crystallization seems to be represented here.

Gneissoid granite near the contact between normal granite and schist (as collected from Franklin Extension Mine) is fine-grained and dark gray. It consists of bands of granite separated by layers rich in biotite. The parallel arrangement of flakes of biotite emphasizes the banded structure. The so-called bands of granite, as seen under the microscope,

are composed of either (1) graphic orthoclase and quartz, or (2) interlocking biotite and relatively finer-grained quartz. These bands average from 1.4 to 2.24 mm. in width. Large crystals of quartz and orthoclase that make the graphic intergrowth may be the same width as the bands. The finer-grained quartz averages from 0.15 to 0.3 mm. across. Veinlets of microcrystalline quartz crosscut or parallel the bands. Besides orthoclase, acid plagioclase (oligoclase-albite) is subordinate. These feldspars are partially kaolinized and sericitized. The ratio between the length and width of the biotite here averages about 3.8 to 1.

There are also belts consisting of aggregates of sericite, magnetite, andalusite, apatite, cordierite, zircon, and chlorite. These belts seem to represent the original schist that suffered thermal metamorphism with the injection of granite. Andalusite in small columnar aggregates parallels the bands.

Granite cuts sharply across the schist in the vicinity of Blue Hill and Brooksville. Megascopically it seems to preserve its normal texture up to its very contact. Microscopic study reveals that here oligoclase, instead of orthoclase and microcline, is the chief feldspar. The biotite shows a reaction rim when it meets plagioclase (Pl. 3, fig. 5). Biotite in the typical granite does not show this feature; in this marginal phase it has straight contours only when it meets quartz. The ratio between the length and width of the biotite flakes averages about 2:1. Near the very contact with the schist, the ratio increases to 5.5:1.

*Age.*—From the contact features discussed above, the conclusion seems justified that granite is the youngest rock mentioned. Smith and Bastin (1907) noticed that in the Vinalhaven district the granite intrudes surface rocks believed to be of Niagaran age. Pebbles of the granite are present in the conglomeratic facies of the Perry formation in the Perry Basin which has been determined to be Upper Devonian from its plant remains (Smith and White, 1905). In the neighboring area of New Brunswick, the granite intrudes the Silurian rocks with notable contact metamorphic effects. It may be concluded, therefore, that the granite intruded sometime during the late Silurian to middle Devonian.

*Some characteristics of the granite intrusive.*—Several characteristics of the granite are noteworthy:

Uniformity of composition.
Distribution parallel to the regional structure.
Intrusion in the Siluro-Devonian period.
Location in the Acadian orogenic belt.
Intrusion cutting the Ellsworth formation into several separate areas. The Silurian beds of the eastern part are also cut by the granite.

All these fit well with some of the important characteristics cited by Daly for a batholith. On the other hand, though noticeable contact metamorphic effects have been mentioned in places, the influence, as a whole, is not intensive. The most severely affected rock is the Ellsworth schist in the Blue Hill district, and it is also in this area that chalcopyrite mineralization is most extensive, suggesting higher temperatures. This is a relatively small area surrounded by the granite. Even here, a little farther from the contact, thermal effects are shown only by the appearance of such minerals as cordierite and andalusite after biotite. It is only in those areas very near the contact that we notice the formation of hornfels from schist.

Near the West Sullivan Mines, rocks adjoining the granite do not show much evidence of contact metamorphism. In areas like Sedgwick, South Brooksville, and east of North Blue Hill, granite cuts the schist sharply without any effect on the schist that is megascopically apparent. Gneissoid granite is limited to almost the very contact with the granite.

From these facts, it is apparent that the granitic mass now exposed was not formed by melting its way through the crust. The melting process would require an enormous amount of localized superheating of the magma, which was evidently not the case here. Furthermore, the remarkable uniformity of the mineral composition rules out the possibility of the wholesale assimilation and, also, metasomatic processes of granitization. This means that the granite must have been introduced mechanically. Field investigation was not detailed enough, however, to give an insight into the mechanics of emplacement.

### ELLSWORTH FORMATION

*General consideration.*—The name Ellsworth schist was first used by Smith and Bastin (1907) to designate the highly metamorphic argillaceous sedimentary rocks abundantly developed around the Blue Hill district. Emmons (1910) later applied the name to the similar rocks in the mining districts of southeastern Maine. Keith (1933a) in his preliminary geologic map of Maine used the name pre-Cambrian gneisses and schists to include the area formerly mapped as Ellsworth schist.

The name so extended includes rocks showing different degrees of metamorphism, from shale to phyllite, hornfels, and schist, with the schist predominating. No positive evidence of pre-Cambrian age has been found. A comparison with the rocks in the eastern part of the area suggests that part of it may be early Paleozoic. At any rate, it is the oldest formation in the district. Therefore, to avoid a controversy that is beyond the scope of this paper, the name Ellsworth formation is used here with no specific age connotation.

*Geographic distribution.*—The Ellsworth is the country rock of the ore deposits of the Blue Hill, Sullivan, Franklin, and Cherryfield districts. Interrupted by the granite outcrops, it occupies three separate areas. The smallest area stretches from Brooklyn to Toddy Pond in the Blue Hill district. Included in this area are several small outcrops within the granite. The second area extends northward from the shore between Blue Hill Bay and Frenchman Bay. The third area lies north of Pleasant Bay.

*Description.*—The schistose part is well developed in the Blue Hill district. The rock is much crumpled, finely foliated, and locally much injected with quartz. It is dark purplish to dark greenish gray. Fresh surface shows alternate light and dark bands, the former being composed essentially of quartz, and the latter rich in micaceous minerals, chiefly biotite. On the weathered surfaces, the more quartzose bands stand out in relief and emphasize the foliation so well illustrated at the hill near the Twin Lead Mine.

At the northern end of Frenchman Bay, phyllite and argillite are predominant.

*Contacts.*—The Ellsworth is in contact with the massive granite. The contact-metamorphosed schist is rather uniform. It is laminated and, though of a purplish tint when seen from a distance of a few feet, consists actually of alternate black and gray layers. These layers are pseudomorphs of the schistose layers in the schist and show the contortions and drag folds characteristic of that rock, but the hornfels has no schistosity and breaks into angular fragments. Its weathered forms are like those of a massive crystalline rock of uniform character.

At the northern end of Frenchman Bay, *lit-par-lit* injection of granitic materials into the sedimentary rocks forms a gneissoid complex as described above. Along the shore of the West Sullivan Mines, no visible effect on the argillite has been found, though the area is very near the granite.

*Structure.*—The schistosity strikes NE.-SW. Locally it may vary from due E.-W. to almost N.-S. Dips vary from low angles to almost vertical.

Around the Blue Hill copper mines, the schistosity strikes from N. 50° W. to N. 85° W., and the dip is about 50° SW. Along the shore of the Blue Hill district, the schistosity strikes about N. 70° W. and dips 45° SW. The general strike of the northern end of Blue Hill Bay indicates that the shore erosion has been influenced by the schistosity.

For a considerable distance between Surry and Ellsworth, the foliation of the schist is nearly horizontal. Along the shore of the West Sullivan Mines, the strike varies from N. 70° W. to almost E.-W., with the angle of dip from northeastward to northward against the granite.

*Petrography.*—Under the microscope, the schistose structure in the biotite schist is especially conspicuous due to the difference in mineral composition of the alternate bands (Pl. 3, fig. 1). The dark bands are composed essentially of biotite with minor andalusite, sericite, cordierite, muscovite, bleached biotite, magnetite, and ilmenite or leucoxene. These minerals increase as the bands become wider.

The light bands are composed essentially of quartz, with small amounts of fresh, striated grains of oligoclase-albite. The quartz grains are interlocking, well oriented parallel to the schistosity, and are clear except for slight replacement by sericite.

The biotite is deep brown suggesting a high content of titanium (Hall, 1941) (further confirmed by the alteration to rutile), with many inclusions of zircon surrounded by pleochroic haloes. The ratio between the length and width of the biotite averages about 4:1. Most of the flakes are larger than the grains of quartz and feldspar. Minerals occurring in minor quantity in the dark bands seem to have been formed at the expense of biotite. The bladed andalusite pseudomorphic after biotite and the close association of magnetite and spinel with biotite all suggest such an origin.

The presence of cordierite and andalusite shows the effect of contact metamorphism. Cordierite is usually altered to sericite. The ilmenite here is transformed commonly into leucoxene, but some of the so-called skeleton crystals are still seen. Pyrite is usually closely associated with biotite. Occasionally it occupies the interstices between quartz grains.

In the more severely metamorphosed areas where schist is transformed to hornfels schistosity is preserved only by dark carbonaceous matter, and remnants of the recrystallized quartz inherited from the schist may still be seen. Under crossed nicols, the schistose structure disappears almost entirely, and minerals like cordierite, andalusite, and muscovite have developed into large crystals without regard for schistosity.

The andalusite includes dusty carbonaceous matter, magnetite, hercynite, and green spinel (pleonaste). These several minerals are possibly all derivatives from the biotite (Pl. 3, fig. 2). Remnants of the biotite, bleached and colorless, are seen at the border of the andalusite.

The cordierite, though in large grains, lacks clear-cut crystal boundary and often intergrows with andalusite and other minerals. Sillimanite, sericite, carbonaceous matter, and possibly some quartz occur as inclu-

sions in the cordierite. Some features, as seen in the slide containing both hornfels and granite, seem worth noting since the contact between hornfels and granite is shown. A sharp contrast between the texture of the two is what might be expected from the one crystallized out from a fluid magma and the other recrystallized in the solid state. Abundant inclusions of different constituents in a mineral grain and the lack of clear-cut crystal boundaries of the minerals are typical of the hornfels.

A distinct contact zone about 0.6 to 1.2 mm. in width, composed chiefly of quartz, feldspars, and myrmekite, is found between the granite and hornfels (Pl. 3, fig. 4). These mineral grains are equigranular and closely interlocking with an average diameter of about 0.06 mm. The explanation for this seems to be similar to that given by several authors for the pegmatites stimulated by assimilation—namely, that the crystallization of igneous rock adjacent to the original country rock proceeds more rapidly and reaches a more advanced stage than the bulk of the rock mass. Differentiation on a small scale is thus advanced; this is evidenced by a biotite belt closely following the contact zone in the granite.

The occurrence of sericite here is interesting. It is generally understood that the volatile constituents in a magma act as fluxes and mineralizers throughout crystallization. Only toward the close of that process, when the temperature has declined greatly, do the same substances begin to exercise a destructive effect upon minerals already crystallized. It is especially at this late stage that the volatile substances can escape into the surrounding rocks. Thermal metamorphism must be assigned in the main to an earlier time, when the temperatures are higher. In general, pneumatolysis and hydrothermal action follow metamorphism and are superposed on it.

This superposition feature is well illustrated by the distribution of sericite (Pl. 3, fig. 3). Andalusite, cordierite, and feldspars in the hornfels are much sericitized. Feldspars in the granite also show less marked alteration to sericite. A sericite veinlet from the granite cuts through the contact zone into the hornfels which had already suffered both regional and thermal metamorphism.

### SILURIAN ROCKS

*Geographic distribution.*—Silurian rocks are fully developed only in the easternmost part of the area (Pl. 1).

*Description.*—The strata consist chiefly of shale grading in places into well-stratified tuff. Volcanics, chiefly rhyolitic flows and tuffs, with some andesite and diabase, are interbedded with the strata and are dominant.

*Contacts.*—The Silurian rocks are separated from the rest of the area by the granite, the intrusive character of which is shown by the decrease in grain toward the contact and by the granite dikes in diabase which intrudes the Silurian. The Silurian rocks near the contact show extensive silicification.

*Structure.*—The structure of the Siluro-Devonian beds in the eastern part of the area has been well described by Bastin and Williams (1914). Essentially, most of the Silurian rocks here are confined to a central fault block which is outlined by two NE.-SW. striking master faults, the so-called Ayers Junction fault and the Lubec fault zone. Within this block, the Silurian rocks have been folded into a broad anticline that pitches gently to the northeast. Their strike shifts from NW. in the northern and central parts to NE. in the southeastern part. Numerous minor faults believed to be a result of tension cut the beds generally in a northeast-southwest direction. These seem to be later than the major features, as they produce offsets in the latter.

*Petrography.*—The shale is composed of very fine-grained quartz, muscovite, chlorite, spinel, sericite, kaolinite, and argillaceous matter. The transition from the rhyolitic tuff to shale is by no means gradual, yet it shows continuous deposition (Pl. 2, fig. 4). Fragments of quartz and feldspars from the tuff have been caught up in the basal part of the shale.

The rhyolitic tuff consists of large angular fragments of orthoclase, quartz, and rock fragments with their interstices now filled with carbonate, sericite, and chlorite. Nearly half the tuff is composed of fragments of feldspars (dominantly orthoclase, with some acid plagioclase), which have an average size of about 1 by 0.5 mm. Large ones may reach twice this size.

The abundance of these large fragments of quartz and feldspars in the tuff is due to the occurrence of these minerals as phenocrysts in the rhyolitic flow. This is a porphyritic, merocrystalline rock with its phenocrysts composed of feldspars, quartz, and lithic fragments in a groundmass of cryptocrystalline quartz and feldspar formed by devitrification.

The rock changes to andesite or diabase when feldspar laths (oligoclase-andesine) occur as phenocrysts and quartz and orthoclase decrease or are absent. Effusive diabase is amygdaloidal, with the amygdules filled with carbonate.

### OTHER INTRUSIVE ROCKS

*Geographic distribution.*—Intrusives other than the granite are of minor quantity and do not concentrate in a particular area. Three separate major areas may be noticed. In the Penobscot Bay area, basic rocks,

mainly diorite, occupy a border zone surrounding a somewhat circular massive granite south of South Penobscot. In the vicinity of Frenchman Bay, diorite forms the country rocks of the Gouldsborough Mine. In the eastern part of the area, diabase intrudes the Silurian rocks.

*Description.*—The diorite ranges from aphanitic to coarse crystalline. The color varies somewhat with the texture; it is dark greenish gray in the finer-grained and mottled white and black in the coarser-grained variety. In the vicinity of Frenchman Bay, diorite is cut by pegmatite veins. The diabase is dark gray, massive, and has common diabasic texture and composition.

*Petrography.*—The coarse-grained diorite is porphyritic, with hornblende in large grains averaging about 1.6 by 0.8 mm. Feldspar laths (andesine) range from 0.2 by 0.1 mm. to 0.8 by 0.3 mm. Quartz is interstitial. The accessory minerals include magnetite, sphene, and apatite.

In the finer-grained variety, hornblende and feldspars are approximately equal in size, though the hornblende is usually somewhat larger. Dark-brown biotite also occurs, sometimes equaling the hornblende.

Near the Gouldsborough Mine, the diorite is cut by pegmatite veins. Under the microscope, the contact between the two may be noted by the sudden decrease of ferromagnesian minerals, increase of quartz content, and the appearance of orthoclase.

The diabase shows its typical ophitic texture with feldspar laths (andesine-labradorite) separated by ferromagnesian minerals, chiefly augite, and its alteration products hornblende and chlorite.

## ORE DEPOSITS

### RELATION OF ORE DEPOSITS TO ROCKS OF THE AREA

*Relation to the granite.*—The ore deposits are almost all clustered around the granite intrusions. In the Blue Hill district, they occur in the Ellsworth schist in the area surrounded by granite, at distances of ¼ to 1 mile from the contact. Those near Frenchman Bay are confined to a belt striking NW.-SE. approximately parallel to the contact, and from ¼ to 2 miles from the granite. Granite crops out ¾ mile northwest of the Cherryfield Mine. The Big Hill Lead Mine and the Lubec Lead Mine are about 2 and 15 miles, respectively, from the granite.

The writer has already indicated that the granite was intruded sometime during the late Silurian to middle Devonian, which is closely related to the Acadian orogenic movement. The ore minerals seem to have been introduced at the same period, since ore veins cut both the Ellsworth formation and the Silurian rocks. As far as known, no metallic deposits have been reported in the near-by areas in the strata younger than the

granite intrusion. Emmons believed that the ore deposits of the Blue Hill district were dynamometamorphosed before the intrusion of granite, but the field, petrographic, and mineragraphic evidences here presented all oppose that theory.

If the ore deposits are genetically associated with the granite intrusion, a certain zonal arrangement around the granite should be expected. This is exactly what is found. Pyritic copper ores occur in the Blue Hill district whose location between two granite masses doubtless accounts for the high-temperature mineralization. Magnetite, arsenopyrite, and pyrrhotite are found exclusively in the deposits near the granite, with pyrite and chalcopyrite particularly abundant. Lead and zinc deposits are located in the Silurian rocks farther from the granite, with sphalerite and galena as chief ore minerals.

*Relation to other rocks of the area.*—Most of the ore deposits are found in the Ellsworth formation, although Lubec and Big Hill lead mines are in the area of Silurian rocks. Diorite forms the country rock of the Gouldsborough Mine.

### ORE DEPOSITS OF BLUE HILL DISTRICT

*Location.*—Blue Hill village lies 14 miles west of Ellsworth. The village owes its name to the near-by Blue Hill which has an elevation of 940 feet and is the highest hill in the area surrounding Penobscot Bay.

The mines are all located south of Blue Hill village. The largest, the Douglas Mine, is about 2 miles southwest on the northwestern shore of Douglas Pond. Owen Lead Prospect lies 1 mile southwest of the Douglas Mine. Twin Lead Mine, Stober Mine, and Blue Hill Mines are all about 1½ miles southwest of the village, and the Granger Mine is about ¾ mile due south of Blue Hill village. All except the Owen Lead Prospect are copper mines.

*History of the development.*—All the mines have lain idle for years and are inaccessible. Only Douglas Mine seems to have ever been worked intensely. Owen Lead Prospect is a small pit 20 feet deep.

The Douglas Mine, opened in 1878, was worked until 1883. It is said to have yielded about $300,000 worth of copper. Other copper mines in the area were operating at about the same time. The Douglas Mine was reopened in 1908 and sampled by interests identified with the Pittsburgh and Montana Company of Butte. Intermittent work was carried on up and through the first World War. It has not been operated since.

*Country rocks.*—Ellsworth schist forms the country rock of all these copper mines. The Owen Lead Prospect covers an area occupied by

the schist intruded by granite dikes. The schist extends roughly in a north-south direction and is almost surrounded by the granite, a factor that must account for the high-temperature type of mineralization as well as the contact metamorphism of the schist.

The granite is quartz-rich, medium-grained, and light gray to light pinkish gray. The schist is highly contorted and consists of alternating layers of dark foliated micaceous minerals and light-colored bands of quartz. Each layer is about 1 to 2 mm. thick, the light bands the wider. The width of the layers, however, is not constant.

In the region of the mines, the strike of the schistosity is prevailingly WNW.-ESE. The dip is 50° SW., though local variation of both can be very great.

*Petrography.*—The results of petrographic studies of the unmineralized rocks of this region have already been presented. Attention is directed here, therefore, to the studies of the mineralized portion in order to show their differences and relations.

Andalusite, cordierite, muscovite, and a possible increase of biotite are the result of contact metamorphism. The development of these minerals, however, is not necessarily accompanied by the introduction of sulphides. Included as it is between masses of granite, the schist in this area has been thermally metamorphosed, but sulphide mineralization is restricted.

Sulphide mineralization in this area is usually associated with the introduction of quartz. The obvious difference between the introduced vein quartz and the recrystallized and reoriented quartz grains in the schist (Pl. 4, fig. 1) affords one of the best evidences against the formation of the ores before regional metamorphism as advocated by Emmons. It must be pointed out, however, that quartz veins do not always indicate the presence of sulphides. Many large quartz veins with no accompanying sulphides occur along the shore near Blue Hill village.

Under the microscope, the original schist bands are shown by alternating layers of micaceous matter, now chiefly sericite, and fine-grained quartz. The recrystallization of the quartz grains is indicated by their elongation in the plane of schistosity, their parallel crystallographic orientation, and their closely interlocking character. Their average diameter is a little over 0.03 mm. The quartz contains minute inclusions of zircon, some apatite, tourmaline, rutile, and sericite.

The foliated minerals in the schist are biotite, sericite, and chlorite. Alteration of biotite releases magnetite and rutile. Scattered among these minerals are altered euhedral pyrite grains which should not be

confused with the pyrite associated with the other sulphides in the quartz veins that were introduced much later. These pyrite grains, though much altered and replaced by biotite, quartz, chlorite, and carbonate, still retain their cubic form. The pyrite grains are shattered and cut by the quartz veins which are the first indication of the introduction of the ore solution.

The vein quartz is coarser than that in the schist and shows no recrystallization. The grain size ranges from 0.24 mm. upward. The quartz contains very few inclusions which are evidently a direct continuation of those in the schist which were incorporated during the replacement of the older by the younger quartz.

The sulphides, other than the pyrite grains, are invariably associated with the vein quartz. These are later than most of the quartz, as evidenced by their interstitial position. Other gangue minerals are chlorite, biotite, sericite, and carbonate. Carbonate is evidently the youngest since it fills the fractures and replaces other minerals.

The sulphide-bearing granite in the Owen Lead Prospect belongs to the marginal phase of the granite as designated in the previous part of this report. It is medium-grained, with oligoclase instead of microline as the dominant feldspar. Otherwise, it has the same mineral composition and the same relative proportion of minerals as the near-by granite. There is also little alteration of the minerals. Some chlorite accompanied by magnetite and rutile is developed from the decomposition of biotite, and the feldspar is slightly kaolinized. Sulphides replace biotite and surround the feldspar crystals.

*Mode of occurrence.*—The ore deposits occur from ¼ to 1 mile from the contact with the main granite mass. The ores in the schist form lenticular masses with more or less parallel alignment and grade into the schist. It is this feature which led Emmons to believe that the ores were formed prior to regional metamorphism. Their abundance only near the contact, however, suggests their genetic relation to the granite. Moreover, in many places, sulphide veins cut across the lamellae of the schist. Particularly good examples of this crosscutting relation occur at the north wall of an abandoned shaft of the Twin Lead Mine. One of these small veins containing quartz, pyrite, and chalcopyrite cuts squarely across the schistosity.

The lodes are said to be 5 to 20 feet wide at depth and to follow the schistosity closely. All the shafts were flooded when this study was made, but sites of the abandoned shafts are arranged more or less parallel to the schistosity of the country rock. Where exposed in an

open cut a few feet northwest of the main shaft, the lode follows the schistosity of the country rock.

A small 20-foot hole, now half flooded, is the only exposed development of the Owen Lead Prospect. At this spot, granite dikes intrude the schist. Owing to rapid weathering of the feldspars in the moist climate, the granite dikes in the wall break down into a granular mass in contrast with the tough, solid schist which it intrudes. The granite is studded with sulphides which appear to be intergrown with the rock minerals.

*Mineragraphy.*—The ore minerals from the copper mines are pyrite, chalcopyrite, pyrrhotite, sphalerite, and magnetite. Secondary minerals include marcasite (after pyrrhotite), covellite (after chalcopyrite), and hydrous iron oxide.

Pyrite is most abundant. Owing to the mutual interference of crystal grains and to replacement, perfect euhedral forms are not common. Straight crystal outlines, however, can often be seen. Single grains are over 2 mm. in diameter.

The pyrite frequently includes quartz grains indicating that pyrite formed later. Magnetite is the only ore mineral older than pyrite, which precedes both chalcopyrite and sphalerite (Pl. 5, fig. 3). It is replaced by pyrrhotite in places. When chalcopyrite is in contact with the boundary of pyrite grains, the contact is usually very sharp and gives no indication of replacement. On the other hand, when chalcopyrite veinlets enter fractures in pyrite, they expand rapidly. The inner part of a crystal seems, therefore, to offer less resistance to replacement than the outer part.

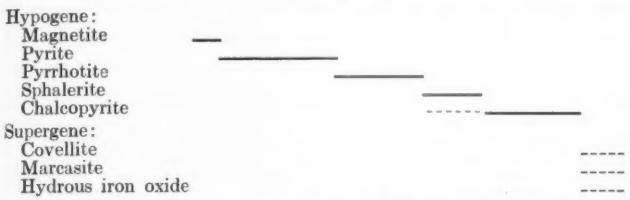
Chalcopyrite is next in abundance. It is younger than pyrite or pyrrhotite since it fills fractures in pyrite and surrounds and replaces grains of pyrite and pyrrhotite. It surrounds sphalerite which in turn contains small globules of chalcopyrite. In places, chalcopyrite has been cut by secondard covellite veinlets.

Thompson (1923) reports that "a polished section showed the occurrence of pentlandite along the microscopic fissures in the pyrrhotite". No pentlandite has been found in the specimens examined by the writer. However, a dimethyl glyoxime test on pyrrhotite powder yielded evidence of the presence of nickel.

Small dark-gray sphalerite grains occur in chalcopyrite, fill fractures in pyrite, and appear to replace pyrite. These sphalerite grains usually contain small globules of chalcopyrite.

Magnetite is rare. When it is found with pyrite, it occurs in the grain of pyrite.

The sequence of ore minerals appears to be as follows:




According to Newhouse and Flaherty (1930), the ore minerals in the granite consist of magnetite, pyrite, pyrrhotite, sphalerite, chalcopyrite, and galena.

*Discussion.*—From the above description, evidence against the predynamo-metamorphic origin of the pyritic copper deposits but in favor of their genetic relation with the granite intrusion may be summarized as follows:

Field evidence:
Abundance of ore deposits near the granite and their absence elsewhere.
Veins of pyrite and chalcopyrite cutting across the lamellae of schist.

Petrographic evidence:
Striking difference between the introduced vein quartz and recrystallized and reoriented quartz grains in the schist.
Chloritization and sericitization of the schist near the vein.
Contact metamorphism of the schist, apparently with increasing Fe,Mg,S,$SiO_2$ content as evidenced by the development of such minerals as cordierite, biotite, and pyrite.

Mineragraphic evidence:
No elongation of crystals of ore minerals parallel to the schistosity.
Veinlets of pyrite and chalcopyrite showing little evidence of crushing and compression.

### GOULDSBOROUGH MINE

*Location.*—The Gouldsborough Mine is located in the Town of Gouldsborough east of Frenchman Bay (Pl. 1).

*Country rocks.*—Diorite is the country rock. Granite crops out about 1 mile away.

*Mode of occurrence.*—Diorite is cut by veins of pegmatite, quartz, quartz and orthoclase, and veinlets of orthoclase. The quartz and orthoclase veins contain sulphides. Sulphides also form veinlets and occur as impregnations in the country rock. Fragments of diorite are sometimes included in the vein quartz.

*Petrography.*—A petrographic description of the diorite and its relation to the pegmatite vein have already been given. Attention is directed here to the discussion of the mineralized area.

Well-banded structure is seen under the microscope. A section (Pl. 4, fig. 2) cut parallel to the surface containing quartz and orthoclase veinlets shows the following features (approximately parallel to each other):

Quartz veinlet about 6 mm. in width composed of coarse-grained vein quartz traversed by minor veinlets of fine-grained quartz.

A belt of altered original country rock about 6 mm. in width composed essentially of very fine-grained epidote, apatite, quartz, chlorite, limonite, sericite, and kaolinite. Mineralization and hydrothermal alteration seem to have so progressed that quartz and apatite are the only original minerals that remain.

A narrow belt of epidote about 0.5 mm. in width with the epidote in large euhedral grains. It seems to have been introduced later than the quartz and orthoclase, since it fills the interstices between quartz grains and surrounds euhedral orthoclase.

Quartz and orthoclase veinlets about 7 mm. in width with orthoclase occupying about ½ to 1/10 of the whole veinlet and more or less confined to the central part. Both orthoclase and quartz are euhedral.

*Wall-rock alterations.*—Chloritization and epidotization are evident as a part of the wall-rock alteration, with some sericitization and albitization. Chloritization is the most pronounced alteration, although it does not extend very far. When a chlorite veinlet passes through hornblende, it may (1) replace hornblende along the cleavage fractures; (2) decolorize the hornblende; and (3) decolorize and wipe out the typical amphibole cleavage. Epidote is usually associated with chlorite in hornblende and biotite. The epidote developed in the country rock is not in large euhedral grains as in the case of the epidote in the vein. Sericite replaces feldspar. When diorite fragments are surrounded by vein quartz, the feldspar in the fragments seems to have been entirely albitized.

*Mineragraphy.*—The ore minerals are galena, chalcopyrite, pyrite, sphalerite, with secondary covellite, anglesite, and hydrous iron oxide.

Galena and chalcopyrite are the two most abundant minerals in the specimens examined. Galena occurs as veinlets filling the fractures in pyrite and also locally replaces it. Galena contains small areas of chalcopyrite and indents both chalcopyrite and sphalerite. It is cut and replaced by secondary covellite along the periphery and along the fractures. The secondary lead mineral, anglesite, replaces galena along the cleavage lines. Both etching with 1:1 $HNO_3$ and the silver chloride test reveal that the galena is argentiferous.

While a small part of the chalcopyrite occurs as small blebs and dots in the sphalerite, most of it occurs independently and encloses sphalerite. In turn, it is cut by veinlets of secondary minerals, especially covellite (Pl. 5, fig. 4).

A few pyrite grains about 0.5 mm. in diameter are cut by veinlets of galena and chalcopyrite.

Sphalerite is not abundant. It contains small inclusions of chalcopyrite which in turn surrounds the sphalerite (Pl. 5, fig. 4). Paragenesis seems to be as follows:

FRANKLIN EXTENSION MINE

*Location.*—The Franklin Extension Mine is located 2 miles west of Franklin Station on Frenchman Bay (Pl. 1).

*Country rocks.*—The Franklin Extension Mine lies near the contact of the Ellsworth formation with the granite. The former consists of schist, phyllite, and interbedded amygdaliodal diabase. The granite near the contact has a gneissoid structure apparently formed by the injection of granitic material along the foliation planes of the schist. Near the contact the schist and the gneissoid granite are so similar on weathered surfaces that it is difficult to differentiate them. The granite also includes fragments of the schist.

*Mode of occurrence.*—Two pits, apparently on the same lode, lie about 300 feet apart. The eastern pit is roughly circular with a diameter of about 50 feet. The western pit is rectangular, about 20 feet long in a NE.-SW. direction, and 10 feet wide. In both places, the ore minerals generally occur as veinlets parallel to the schistosity, but they occasionally cut across it. They occur also as impregnations in the Ellsworth formation.

*Petrography.*—Petrographic descriptions of the granite, gneissoid granite, and schist have been given previously. It has been mentioned that there are belts of gneissoid granite composed of aggregates of sericite, magnetite, andalusite, apatite, cordierite, zircon, and chlorite, apparently representing the original schist that has suffered contact metamorphism due to the injection of granite. On the other hand, small granitic veins injected into the Ellsworth formation may have no effect at all, with the possible exception of an increase of biotite and the development of feldspars.

The phyllite which contains sulphide veinlets, is well foliated as a result of the parallel arrangement of sericite, chlorite, quartz, and carbonaceous matter. Quartz occurs as enclosures. The individual quartz

grains average about 0.05 mm. in diameter, but those occurring as enclosures may be 1.6 mm. long and 0.8 mm. wide, with the longer direction parallel to the schistosity. The sulphide veinlets, chiefly pyrite, parallel the schistosity but occasionally cut across it. Closely associated with the pyritic veinlets is carbonate which also fills the fractures in other minerals.

The amygdaloidal diabase is composed of feldspar laths (andesine-labradorite) and chlorite, with amygdules consisting dominantly of quartz, with epidote and pyrite. The feldspar laths average about 0.2 by 0.04 mm. in size. They are partially sericitized and kaolinized.

*Mineragraphy.*—Ore minerals are arsenopyrite, pyrite, pyrrhotite, sphalerite and chalcopyrite.

The arsenopyrite is galena-white, distinctly anisotropic, with slight absorption. Polarization color ranges from light gray, light pinkish gray, to pale grayish yellow. It exhibits beautiful diamond-shaped form. Single grains range from 0.1 mm. (shortest diagonal) and 0.14 (longest diagonal) to 0.7 mm. (shortest diagonal) and 1.2 mm. (longest diagonal).

Short (1940) states that arsenopyrite is harder than pyrite. In the specimens examined, arsenopyrite is always weaker than pyrite, as the "Becke line" always goes toward arsenopyrite when the microscope tube is raised. While arsenopyrite is definitely older than pyrrhotite and chalcopyrite which fill the interstices between arsenopyrite grains (Pl. 5, fig. 5), its relation with pyrite is not so clear. It seems, however, to have crystallized out earlier than pyrite, as it is surrounded by pyrite grains in many places.

Pyrite is the most abundant mineral in the specimens examined, closely followed by arsenopyrite and pyrrhotite. It is always cubic ranging from 0.2 to 1 mm. in diameter. Schneiderhöhn (1931) states that anisotropic pyrite contains arsenic. In the specimens examined, though arsenopyrite and pyrite are closely associated, the latter is always isotropic.

Pyrrhotite occurs as veinlets or as filling in the fractures of pyrite and arsenopyrite. Chalcopyrite is rare. When found, it is always closely following pyrrhotite as filling in the fractures of pyrite and arsenopyrite. Sphalerite is seen only in a few places and seems to replace pyrrhotite, arsenopyrite, and pyrite. Its relation with chalcopyrite is not clear, as the two do not occur together in the specimens examined.

Paragenesis seems to be as follows:

### COPPEROPOLIS MINE

*Location.*—The Copperopolis Mine is near Egypt on Frenchman Bay about 4 miles west of Franklin Station (Pl. 1).

*Country rocks.*—Chlorite schist is the country rock of the mine. The schistosity here strikes N.20° W. and dips 25° S.W.

*Mode of occurrence.*—Three pits have been sunk in the contorted schist which carries seams of ore. The veinlets seem to have been introduced along the fractures. The lode is said to be a fissure vein in line with that of the Sullivan Mine which passes northwest under the Bay.

*Petrography.*—The slides show typical schistose structure formed through the parallel arrangement of recrystallized quartz, chlorite, and sericite. Scattered among these laminae are small eyes of quartz, feldspars, muscovite, and tourmaline. These eyes, especially those of quartz, contain inclusions of tourmaline with definite orientation, though not necessarily parallel to the schistosity.

*Mineragraphy.*—Cubic pyrite is replaced by chalcopyrite which contains remnants of pyrite from the large crystals. Chalcopyrite also contains small areas of sphalerite which in turn contains small globules of chalcopyrite. Galena replaces chalcopyrite in places. The latter is also traversed by several veinlets of secondary covellite and hydrous iron oxide. The sequence seems to be as follows:

### WEST SULLIVAN MINES

*Location.*—The West Sullivan Mines are located all along the east shore of Frenchman Bay near Sullivan (Pl. 1).

*Country rocks.*—The ores are in the Ellsworth formation which consists of alternating metamorphosed argillaceous and arenaceous sedimentary rocks. Near the Sullivan-Waukeag Bridge, the rocks strike N. 70° W. Along the shore where the mines are located, the strike is prevailingly E.-W. to NW.-SE., and the dip is about 20° N. toward the granite which crops out at a short distance to the east. Near the town of Sullivan, where the granite is in contact with the sedimentary rocks, gneissoid granite is found.

*Mode of occurrence.*—During the time of visit, all the shafts were inaccessible. It is reported (Kempton, 1879) that all the shafts are on the same lode which is a fissure vein running in a NW.-SE. direction approximately parallel to the contact with the granite. The vein also dips at a low angle toward the granite. The vein is occasionally cut by dikes of diabase.

To judge from the dump, quartz is the most important gangue mineral. It is sometimes massive; sometimes it shows good comb structure with acicular crystals pointing to the center of a druse and occasionally it includes fragments of the country rock. Pyrite is the most abundant sulphide associated with quartz. This quartz-pyrite combination suggests formation at high temperature. Sulphides occur in the veins of quartz and pyrite, as stringers, and as impregnations in the country rock.

*Petrography.*—The alternating argillaceous and arenaceous metamorphics as seen under the microscope are composed of quartzitic beds alternating with beds composed of sericite, fine-grained quartz, chlorite, epidote, apatite, magnetite, pyrite, and dusty argillaceous matter. The quartz grains are closely interlocking but have no definite orientation. The quartz grains in the quartzitic beds average about 0.5 mm. in diameter; those in the argillaceous beds are about 0.15 mm.

The pyrite in the vein is closely associated with sericite. The pyrite mineralization followed the introduction of most of the quartz, since the pyrite-sericite veinlets crosscut the quartz. Chlorite occurs interstitially between quartz grains. Carbonate occurs occasionally in the interstitial part and is probably one of the latest minerals.

*Mineragraphy.*—Kempton (1879) reports that the ore minerals here include native silver and such silver-bearing minerals as stromeyerite, stephanite, and argentite. The shafts are now all under water, and good specimens are difficult to obtain. None of the specimens examined by the writer yielded any silver-bearing minerals.

Pyrite is the most abundant mineral. It is commonly in euhedral grains about 0.06 mm. in diameter. Large grains may reach over 0.6 mm. in diameter and are usually broken and interfere with each other.

Arsenopyrite is frequently found in rectangular forms, sometimes triangular or diamond-shaped. The largest ones observed are about 0.02 by 0.05 mm. in size. Hence, when it occurs with pyrite, its presence may be overlooked due to the slight difference in color. It is easily noticeable when these small grains are surrounded by dark-gray sphalerite. While it is definitely earlier than sphalerite, its age relation to pyrite is not so clear. It seems, however, to have crystallized earlier than pyrite, since it is often surrounded by pyrite grains.

Sphalerite in small grains contains dots of chalcopyrite, and chalcopyrite also surrounds the sphalerite areas.

Paragenesis seems to be as follows:

BIG HILL LEAD MINE

*Location.*—Big Hill, with an elevation of 300 feet above sea level, is located about 2 miles northwest of West Pembroke in the northwestern portion of the Eastport Quadrangle.

*History of the development.*—During 1906, considerable work was done, and shallow openings were made at a dozen places on the hillside. The property has changed hands several times since. The prospects now lie idle. The largest open cut visited is located at the southwestern slope of the Big Hill.

*Country rocks.*—The prospects are in the Silurian Pembroke formation which consists of alternating beds of shale and rhyolitic to diabasic volcanics.

At the western wall of the open cut, the following section is exposed (from S. 10° W. to N. 10° E.) (Pl. 2, figs. 1-3):

| | Inches |
|---|---|
| Shale: Striking N.70°E., standing almost vertical at the surface. Partially covered. | |
| Diabase: Containing fragments of phyllite near its contact. Cut through by a vertical ore veinlet. | 52 |
| Shale: | 26-30 |
| Rhyolite: | 26 |
| Shale: | 13 |
| Dacite with diabase at top forming an unconformable contact with shale. | |

*Mode of occurrence.*—The ore occurs as veins filling fractured zones, cutting shales and traps, as disseminations in the country rock, and as amygdaloidal filling in the vesicular lava. Most of the veins are steeply inclined.

*Petrography.*—The shale is composed essentially of very fine-grained quartz, sericite, and argillaceous matter. The stratification is apparent from the dominance of argillaceous matter and fine-grained quartz in one layer and comparatively coarser-grained quartz with less argillaceous matter in other layers.

The rhyolite is a merocrystalline, porphyritic rock with phenocrysts of quartz, orthoclase, and lithic fragments in a groundmass of devitrified

glass. The large quartz phenocrysts reach about 0.16 mm. in diameter, while the orthoclase phenocrysts are about 1.10 by 0.30 mm.

The dacite is a merocrystalline, porphyritic rock with phenocrysts of quartz, orthoclase, and acid plagioclase in a groundmass of microcrystalline feldspar laths and cryptocrystalline devitrified glass. The rare quartz phenocrysts are never so large as the feldspars which average about 0.25 by 1.00 mm.

Andesite is the country rock of a small 7-foot pit on the top of Big Hill. The rock is composed essentially of feldspar laths (oligoclase-andesine) which have an average size of 0.3 by 0.1 mm. Some may reach twice this size.

The intrusive diabase outcrops nearly 1 mile southeast of the mine. It has a typical ophitic texture. The andesine laths are separated by ferromagnesian minerals, chiefly augite. The diabase near the open cut is so altered that none of the original constituents remain unchanged.

*Wall-rock alterations.*—Chloritization has affected nearly all wall rocks, particularly the more basic. In one dacite porphyry examined, nearly all the feldspars are replaced by chlorite. Feldspars have also been more or less sericitized. In the extreme case, nearly 90 per cent of the feldspars have been replaced by fine shreds of sericite. Pyritization is also evident. Pyrite grains are seen in all the rocks examined. In the shale, pyrite occurs roughly parallel to the stratification. Carbonate occurs in ore-bearing veins, as veinlets cutting through the country rocks, and also replaces feldspars in some places.

*Gangue minerals.*—Quartz and carbonate are the most important gangue minerals. Sericite is found in rock fragments which are caught up in the introduced veins (Pl. 4, fig. 4). The grains of quartz range from microcrystalline aggregates to coarse euhedral crystals. Quartz sometimes exhibits comb structure; more often, it is flamboyant. In the main, vein quartz (Pl. 4, fig. 5) was introduced before the ore minerals. Some seems to have come in later since it forms a rim around the ore minerals. Narrow chlorite veinlets occasionally cut vein quartz but they are replaced by carbonate, which also replaces both quartz and ore minerals.

*Ore minerals.*—The ore minerals are pyrite, chalcopyrite, pyrrhotite, sphalerite, and argentiferous galena. Sphalerite and galena are most abundant. Pyrite in small grains is surrounded by sphalerite. Pyrrhotite and chalcopyrite occur in small inclusions in the sphalerite. Some small globules in the sphalerite contain both chalcopyrite and pyrrhotite, and the pyrrhotite seems to be replaced by chalcopyrite. Sphalerite also

occurs as small areas in the galena, which replaces sphalerite. The galena itself is replaced by secondary anglesite along the cleavage traces. The sequence of the hypogene minerals seems to be as follows:

Pyrite ——
Sphalerite (with chalcopyrite, pyrrhotite) ————
Galena ————

LUBEC LEAD MINE

*Location.*—About 7½ miles southwest from Lubec, the easternmost town of the United States, the mine covers a high cliff between South Bay and Bassett's Creek. The shore along which the prospects are located is everywhere steep. Deep cuts occur along the shore cliff.

*History of the development.*—The mine is said to be the first of the kind discovered in Maine. The deposits were discovered in 1828. It has laid idle for over half a century, and the deep workings are submerged.

*Country rocks.*—The country rocks of the mine are vesicular lava, shale, and diabase. Diabase outcrops over a wide area south of Lubec. Near the open cuts, it intrudes the shale so intricately as to form alternating beds with the shale. The contact plane, parallel to bedding of the shale, which, near the contact, has an E.-W. strike, dips about 40° S.

*Mode of occurrence.*—The deposits are closely spaced intersecting fissure veins. Most of the veins strike approximately E.-W. parallel to the strike of the shale and dip steeply toward the south. Some of them, however, cut across these veins. Quartz and carbonate are the important gangue minerals of the vein which sometimes contains fragments of country rocks.

*Petrography.*—Chloritization and sericitization are both evident. Diabase shows typical ophitic texture with partially sericitized andesine laths separated by chlorite, carbonate, and iron minerals. No pyroxene grains were seen. Both chlorite and sericite occur as veinlets cutting the specimens.

The vesicular lava is an amygdaloidal, porphyritic, and merocrystalline rock. Amygdules are filled with carbonate and chlorite. Both also occur as veinlets and replace other minerals. The groundmass is composed of andesine laths and devitrified glass. The phenocrysts are mostly orthoclase. The presence of this mineral is probably explained by the fact that rhyolite outcrops not far away.

*Mineragraphy.*—Galena is the most abundant ore mineral. Both the etching test with 1:1 $HNO_3$ and the microchemical chloride test indicate that it is argentiferous (Pl. 3, fig. 6). Sphalerite contains dots of chalcopyrite, which also occurs independently and sometimes is surrounded

by galena. Galena replaces sphalerite. Pyrite grains, about 0.4 to 0.8 mm. across, are replaced by both sphalerite and galena. The sequence seems to be as follows:

Pyrite ——
Sphalerite ———
Chalcopyrite -----———
Galena ———

### CHERRYFIELD MINE

*Location.*—The mine is located about half a mile east of the town of Cherryfield.

*History of the development.*—The mine was first opened in 1878, was operated intermittently for some time, but is now idle.

*Country rocks.*—The country rocks are shale, more or less metamorphosed, and schistose diabase which is cut by dikes of diorite. Granite crops out ¾ mile northwest of the mine.

The shale and the schistose diabase are in conformable contact. The schistosity of the diabase and the strike of the shale are E.-W., and dips are almost vertical. The schistose diabase is cut by two parallel diorite dikes about 40 feet apart which strike N. 15° E. to N. 20° E. and dip about 75° SE.

*Mode of occurrence.*—To judge from the dump, the deposit is a fissure vein which exhibits banded structure. The ore minerals are in the center and are flanked by gangue quartz. It contains fragments of country rock.

*Gangue minerals.*—Chlorite is developed in the wall rock. Sericite is present but not conspicuous. The gangue minerals consist of quartz, carbonate, and feldspars.

The quartz grains in the larger veins show a comb structure. In the central part, where the ore minerals are abundant, the quartz grains are finer and closely interlocking. The feldspars are microcline-perthite and albite. Carbonate (dolomite) was apparently one of the latest minerals introduced. It replaces the other minerals. In some veins carbonate makes up almost 90 per cent of the gangue and replaces quartz. Sericite and chlorite sometimes occur as small veinlets cutting the vein quartz (Pl. 4, fig. 3).

*Ore minerals.*—Ore minerals are pyrite, pyrrhotite, chalcopyrite, sphalerite, and galena.

Pyrite grains are scarce. They are small (0.05 mm. to 0.5 mm. in diameter), cubic, and surrounded by sphalerite. Pyrite extends into the country rock as impregnations.

Galena contains small areas of sphalerite and also occurs as veinlets cutting deep into sphalerite (Pl. 5, fig. 6). It is argentiferous.

Sphalerite is abundant in the specimens examined. It usually contains small inclusions of chalcopyrite and pyrrhotite. Their mutual relations are discussed more fully in a later section of this paper. Paragenesis seems to be as follows:

Pyrite
Sphalerite (with pyrrhotite, chalcopyrite)
Galena

### DISTRIBUTION AND SEQUENCE OF ORE MINERALS IN RELATION TO GRANITE

*Distribution.*—The relative abundance of some ore minerals in some mines and their total absence in others is dependent upon the distance of the mine from the granite.

Pyrite is the most persistent ore mineral and is found in all the mines discussed. It is particularly abundant, however, in the Blue Hill district, where chalcopyrite, pyrite, and pyrrhotite are the three chief ore minerals. Magnetite also occurs here. The location of the district surrounded by two granite masses accounts for this high-temperature mineralization.

Ore specimens from the mines along Frenchman Bay, where granite occurs on one side only, show galena as a minor accessory. Chalcopyrite and pyrite remain the chief ore minerals. In the mines very near the contact, such as the Franklin Extension Mine, pyrrhotite and arsenopyrite are abundant.

Toward the eastern part of the area, where the mines in the Silurian rocks are located at a considerable distance from the granite, sphalerite and galena are the chief ore minerals. Pyrrhotite occurs only as inclusions in the sphalerite, and pyrite and chalcopyrite are in very subordinate quantity. In the Big Hill Mine, about 2 miles away from the granite, sphalerite seems to be more abundant than galena. In the Lubec Mine, about 15 miles from the granite, the two appear in equal quantities.

The zonal distribution is thus evident. Copper ores occur in the areas near and surrounded by granite. Zinc and lead deposits occur farther from the granite.

*Sequence.*—It is to be expected, therefore, that not all the ore minerals mentioned will be found in the specimens from any one mine. Yet it is significant that the general sequence as listed below holds good:

Magnetite
Arsenopyrite
Pyrite
Pyrrhotite
Sphalerite
Chalcopyrite
Galena

Magnetite and arsenopyrite have not been found together. Their sequence as listed is tentative under the assumption that their ordinary relationship holds true here. If this is the case, then the paragenesis does not differ to any great extent from that of similar kinds of deposits formed at a relatively high temperature and fits well with that given by Lindgren. It accords, almost entirely, with the normal sequence for hypo- and mesothermal deposits as given by Brady (1940) in his theory of the relation between the progressive change in the weight percentage of the anion radical and succession of minerals. Bandy, however, is not certain about the position of sphalerite. He places sphalerite (with a question mark) as later than chalcopyrite. The present suite of specimens shows that most of the chalcopyrite is later than sphalerite.

*Chalcopyrite-bearing sphalerite and pyrrhotite-bearing sphalerite.*—Nearly all the sphalerite observed contains either chalcopyrite or pyrrhotite, occasionally both, as inclusions. The inclusions are in dots, small globules, or dashes. They may be well oriented or distributed at random. While these features are not uncommon, opinions regarding their origin have been contradictory. Some consider the chalcopyrite inclusions in sphalerite a result of replacement of sphalerite by chalcopyrite (Guild, 1914; Teas, 1918; Newhouse, 1926; Ross, 1935); others incline to interpret the feature as a result of breakdown of a solid solution (Schneiderhöhn, 1922b; Van der Veen, 1925). The Subcommittee of the National Research Council cannot decide on the issue and declares: "Both interpretations may be correct in different instances, and no criteria have been formulated for distinguishing between these possible origins." (Bastin, *et al.*, 1931.)

The present suite of specimens seems to indicate replacement of sphalerite by chalcopyrite rather than exsolution. In the case of exsolution, the mineral present as inclusions is generally absent outside of such intergrowth. Here the sphalerite is often surrounded by large areas of chalcopyrite in specimens from the areas near the granite. In specimens from the lead mines, where little chalcopyrite is found outside the sphalerite, the sphalerite contains also very few chalcopyrite inclusions. Moreover, the chalcopyrite inclusions are always distributed at random in the sphalerite (Pl. 5, fig. 1).

On the other hand, if these inclusions of chalcopyrite are considered to have formed by replacement, the question may be asked why there are no chalcopyrite veinlets connecting the inclusions in the sphalerite in spite of the fact that small areas of sphalerite containing chalcopyrite inclusions are in turn surrounded by large areas of chalcopyrite.

The pyrrhotite inclusions in the sphalerite appear, however, to have formed definitely as a result of exsolution, as: (1) They have definite orientation (Pl. 5, fig. 2) indicating control by the crystallographic axes of sphalerite; (2) the included mineral forms disconnected units rather than the continuous network characteristic of replacement along cleavage; (3) there is no enlargement where two directions of pyrrhotite inclusions join or cross; and (4) in specimens in which sphalerite contains pyrrhotite inclusions there is no "independent" pyrrhotite (*i.e.*, pyrrhotite not occurring as inclusions); where there is abundant "independent" pyrrhotite, sphalerite contains inclusions of chalcopyrite only. This seems to suggest a difference in temperature during formation. More sections might, however, reveal the feature to be purely accidental.

### SUPERFICIAL ALTERATION

Alteration of these ore deposits is usually confined to a thin surface coating of limonite. Oxidation is slight, and in most places sulphides appear almost at the surface. Polished surfaces of ores show the development of some secondary minerals along fractures or cleavage traces, such as covellite after chalcopyrite, anglesite after galena, and hydrous iron oxide as veinlets cutting across primary minerals. These are all very minor, however, especially since the specimens come from the prospects which have been developed on or very near the surface.

Several factors account for the slight degree of superficial alteration: the location of the area in the cold temperate zone, the low relief of rolling New England upland, the lack of great fracturing of the deposits, and the relatively recent glaciation.

### ORIGIN

The discussion so far given indicates that there is an obvious relation between the ore deposits and the intrusion of the granite. The evidence is as follows:

The clustering of the ore deposits around the granite intrusion;
The close time relation between the formation of veins and the intrusion of granite;
The activity that the granitic magma has shown in: contact metamorphism in the surrounding rocks, the injection of granitic material into the foliation planes of surrounding rocks near the contact, the granite dikes cutting through the schist, and pegmatite and orthoclase-quartz veins intruding diorite;
The rude zonal arrangement of mineral deposits according to their relative distance from the granite;
The existence of mineralogically similar deposits in the granite itself;
The common association of lodes of this type with acidic or granitic intrusive masses.

It seems reasonable to assume that the ore deposits have been formed by fluids that were given off from the granite intrusion. The sequence of events may be inferred to have been as follows:

(1) Deposition of earlier Paleozoic sediments with intensive volcanic eruptions during the Silurian.

(2) Intrusion of the granitic magma.

(3) Contact metamorphism of the surrounding rocks especially in areas included between granitic masses.

(4) Progressive cooling of the magma.

(5) Rise of vapor pressure.

(6) Fractures produced in the surrounding rocks as the result of cooling after heating by the intrusive magma, and in the marginal granite itself.

(7) Expulsion of metalliferous fluids.

(8) Fluids introduced into the surrounding rocks filling fractures and replacing some of the rocks.

(9) Hydrothermal alteration of country rocks.

(10) Formation of vein materials, introduced quartz usually deposited first, followed by sulphides which are accompanied and followed by gangue minerals.

(11) Superficial change: glaciation and oxidation.

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## EXPLANATION OF PLATES 2-5

### PLATE 2

#### BIG HILL LEAD MINE

Figures 1-3. Geologic Section.
Western wall of an open cut of Big Hill Lead Mine.

Figure 4. Transition from Rhyolitic Tuff to Shale.
Specimen from West Pembroke. Large grains of orthoclase (sericitized and kaolinized) from the tuffaceous part in contact with the shale which is composed of loosely distributed grains of quartz (white), sericite (slender), and argillaceous materials (black). Ordinary light.

FIGURE 2

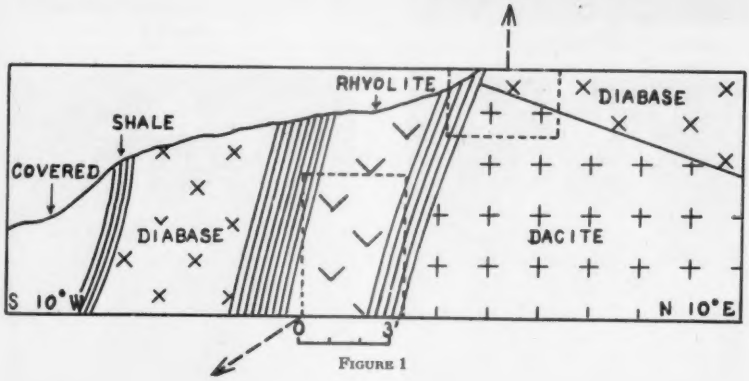



FIGURE 1

FIGURE 3

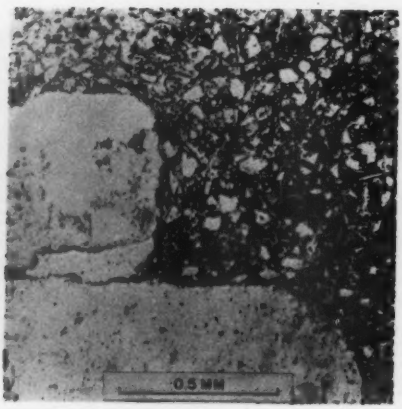



FIGURE 4

BLUE HILL LEAD MINE

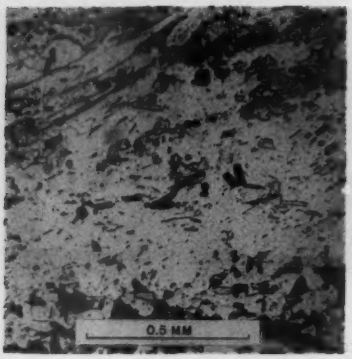



FIGURE 1

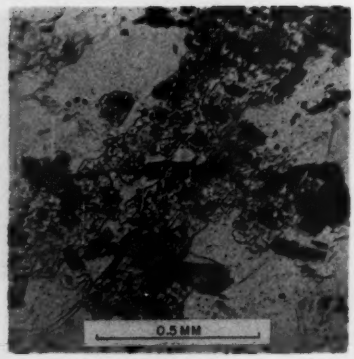



FIGURE 2

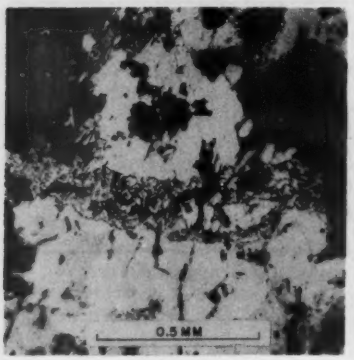



FIGURE 3

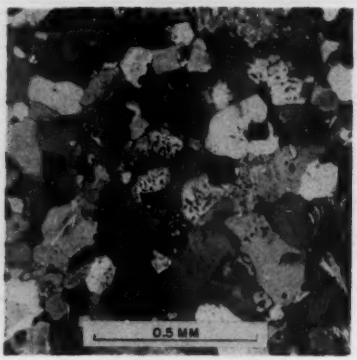



FIGURE 4

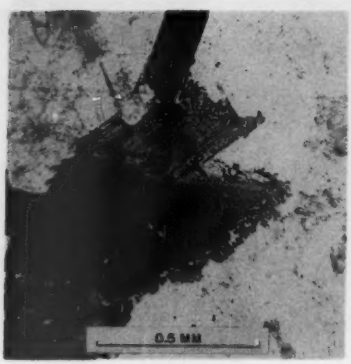



FIGURE 5

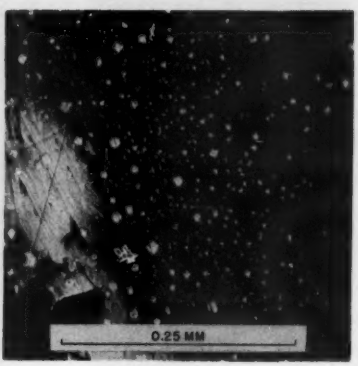



FIGURE 6

THERMAL METAMORPHIC EFFECTS AND ARGENTIFEROUS GALENA

## PLATE 3

### THERMAL METAMORPHIC EFFECTS AND ARGENTIFEROUS GALENA

Figure 1. Quartz-biotite Schist.
Showing the schistose structure formed chiefly by layers of quartz (white) and biotite (dark). In the upper left corner are muscovite flakes (white) with remnants of biotite. In the upper right corner are columnar aggregates of andalusite pseudomorphic after biotite. Quartz with small inclusions of zircon. Black grains in the upper part are magnetite formed after biotite. Ordinary light.

Figure 2. Hornfels.
Showing the development of andalusite after biotite (dark gray) with release of small grains of magnetite, hercynite, and pleonaste (all black). The surrounding area is chiefly cordierite. Ordinary light.

Figure 3. Sericite Veinlet across Cordierite.
The small black area in the cordierite above the veinlet is andalusite. Polarized light.

Figure 4. Contact between Granite and Hornfels.
Polarized light.

Figure 5. Biotite in Marginal Phase of Granite.
Notice the reaction rim of biotite against feldspars. In the upper central part, the contour is straight when the biotite is in contact with the quartz. Ordinary light.

Figure 6. Argentiferous Galena etched with 1:1 $HNO_3$.
Specimen from Lubec Lead Mine. Oil immersion.

## PLATE 4

### GANGUE MINERALS AND WALL-ROCK ALTERATIONS

Figure 1. Introduced Quartz Vein in the Mineralized Schist.
Specimen from Blue Hill district. The quartz vein here shows cracks perpendicular to the schistosity. This feature is not seen in the larger veins. The parallel schistose structure is chiefly made of fine-grained quartz and sericite. Notice the pyrite grains in the upper left corner shattered by quartz veins. There are two generations of pyrite. Most of the pyrite is associated with other sulphides that were introduced later than most of the vein quartz, and is not shown here. Another narrow black area along the contact with vein quartz in the lower left is biotite, now converted to rutile. Ordinary light.

Figure 2. Banded Quartz-orthoclase veinlet in Diorite.
Specimen from Gouldsborough Mine. Starting from the left: country rock (black in the lower left corner); epidote (dark gray with crystal outline); quartz (white); chlorite (a narrow belt); quartz (white) and orthoclase (black); then same again but in reversed order (not shown here). Ordinary light.

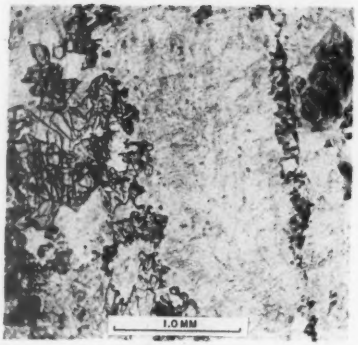
Figure 3. Veinlet of Fine-grained Quartz and Sericite.
Cutting through the coarser-grained vein quartz. Polarized light.

Figure 4. Vein Material.
Specimen from Big Hill Lead Mine. Replacement of a fragment of country rock by sericite (upper central part: country rock, gray; sericite, white and fibrous) which is in turn replaced by carbonate. Carbonate also replaces sphalerite (black, in small triangular form in the lower left). Sericite is found only along the contact with country rock. Ordinary light.

Figure 5. Vein Quartz.
Specimen from Big Hill Lead Mine. Small amount of sphalerite is seen at the lower left, paralleling and also cutting across the quartz. Polarized light.

Figure 6. Sulphide-bearing Bands (dark gray) in the Ellsworth Schist (light gray).
Near Stober Mine, Blue Hill.

FIGURE 1

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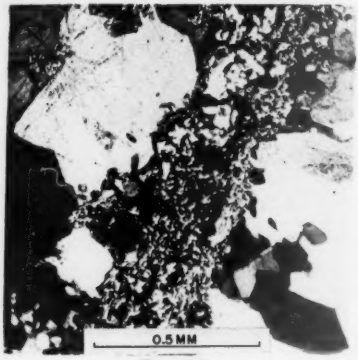


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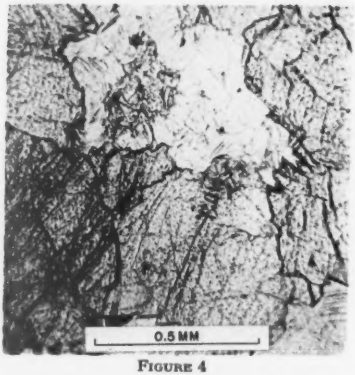



FIGURE 4

FIGURE 5

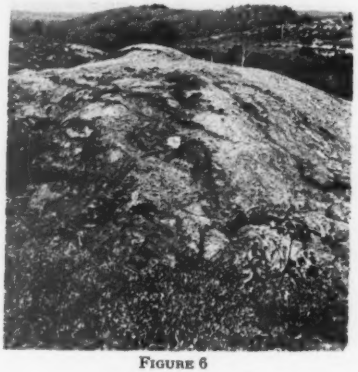


FIGURE 6

GANGUE MINERALS AND WALL-ROCK ALTERATIONS

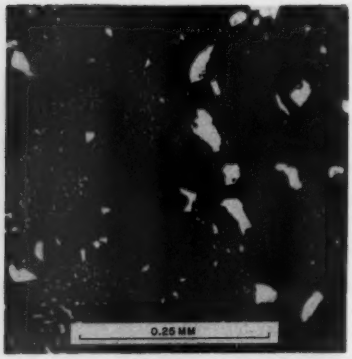
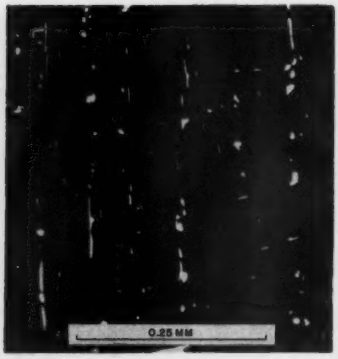



FIGURE 1

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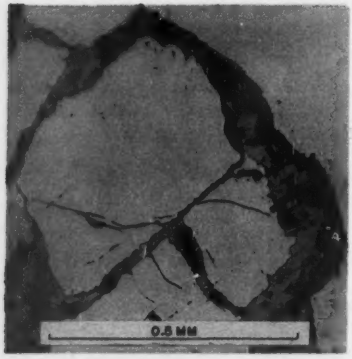


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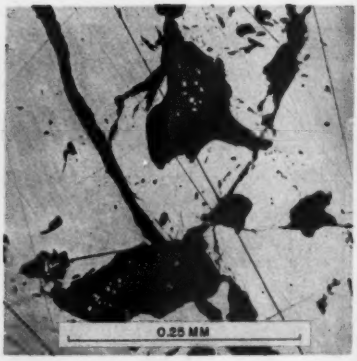


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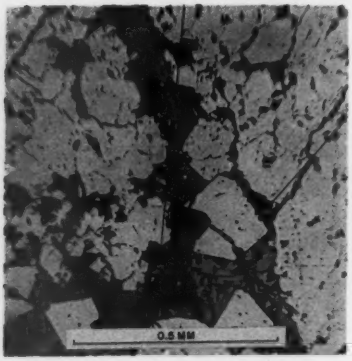


FIGURE 5

FIGURE 6

PARAGENESIS OF ORES

## PLATE 5

### PARAGENESIS OF ORES

Figure 1. Chalcopyrite Inclusions (white, large and small) in Sphalerite.
Specimen from West Sullivan Mines. Oil immersion.

Figure 2. Oriented Pyrrhotite Inclusions (white) in Sphalerite.
Specimen from Cherryfield Mine. Considered as a feature of exsolution. Oil immersion.

Figure 3. Chalcopyrite Veinlets (gray) cut through Pyrite.
Specimen from Blue Hill district. The veinlets contain small areas of sphalerite in places. Oil immersion.

Figure 4. Sphalerite in Chalcopyrite.
Specimen from Gouldsborough Mine. Sphalerite (two large dark-gray areas) with chalcopyrite inclusions is surrounded by chalcopyrite, which is in turn cut by covellite veinlet (in NW.-SE. direction). Oil immersion.

Figure 5. Specimen from Franklin Extension Mine.
Showing the mutual relationship between pyrite (white, in large euhedral grains), arsenopyrite (white, in upper right and left), pyrrhotite (gray, in the interstices between arsenopyrite grains), and sphalerite (dark gray in the upper central part). Sphalerite and pyrite seem to form a veinlet through arsenopyrite. Oil immersion.

Figure 6. Specimen from Cherryfield Mine.
Pyrrhotite-bearing (white dots) sphalerite cut by vein of galena (white), which is in turn replaced by secondary anglesite along fractures. Oil immersion.

VIEW OF WAITE HILLS

Showing bare exposed bedrock surface, Waite area

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# STRUCTURAL FEATURES OF THE KEEWATIN VOLCANIC ROCKS OF WESTERN QUEBEC

BY M. E. WILSON

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## ABSTRACT

The Archean basal complex of the Noranda district, Quebec, consists chiefly of alternating belts of Keewatin rhyolite and andesite with subordinate pyroclastic breccia and stratified tuff. These rocks have been folded with increasing intensity from north to south into eastward-pitching anticlines and synclines. To the north, where the folding is moderate, the lava belts trend sinuously from north to south and dip 45 degrees or less to the east, whereas to the south, where the folding is intense, they trend east-west and are vertical or nearly vertical.

The structure of the lavas has been determined from the following data: (1) the strike and dip of interbedded stratified tuff and chert; (2) the attitude of columnar jointing; (3) the flow contacts within the flow belts; (4) the flow belts; and (5) tops determined by the presence of: (a) zones of breccia along the upper margins of individual flows, (b) the rounded tops and flat bottoms of pillows of the "bun" type, (c) the rounded tops and V-shaped bottoms of pillows of the "balloon" type, and (d) cross-bedding in stratified tuff.

Within an area of 45 square miles extending from Noranda northward an uninterrupted succession of volcanic rocks, whose average total thickness is about 25,000 feet, has been determined. It is proposed to call this succession the Abitibi series.

## INTRODUCTION

When the name Archean was proposed by J. D. Dana for the pre-Cambrian and was later redefined by the United States Geological Survey to include only early pre-Cambrian formations, it was thought that the Archean was largely composed of crystalline schists and gneisses of igneous origin "radically different" in "character" from later rocks and so complex structurally that the methods used in working out the structure of later formations were inapplicable to it. It is now known, however, that the Archean of the Canadian Shield includes a large proportion of lavas and sediments similar to younger rocks and that where they are reasonably well exposed their structure can be determined. This paper is a description of how the structure of the Archean (Keewatin) volcanic rocks of the Noranda district, in western Quebec, was ascertained.

The territory originally investigated (Fig. 1) extends with a width of 4 to 5 miles for 10 miles northward from Noranda and has an area of about 45 square miles. The well-exposed bedrock surface made this an especially favorable district for study (Pl. 1). Its geology is shown on five contoured base maps having a scale of 800 feet to 1 inch published by the Canadian Geological Survey. These are known as the Rouyn, Amulet, Dufault, Waite, and Newbec areas. Additional areas underlain by Keewatin volcanic rocks are being mapped by the writer in the district south of Noranda. In the region to the east where the rocks are not so well exposed, detailed mapping has been carried on in recent years by H. C. Gunning and J. W. Ambrose. An area 5 miles west of Noranda was also investigated in 1940 by E. D. Kindle. The results of all these studies are being published on maps having a scale of 1500 feet to 1 inch.

## GENERAL GEOLOGY

Except for the nearly flat-lying strata of the (Huronian) Cobalt series which form the Kekeko Hills southwest of Noranda and some or all of the numerous dikes of diabase that intersect the rocks of the region, the bedrock formations of western Quebec belong to the Archean basal complex. The lavas and sediments of this complex are usually divided into two groups, the Keewatin and Timiskaming. It has been customary

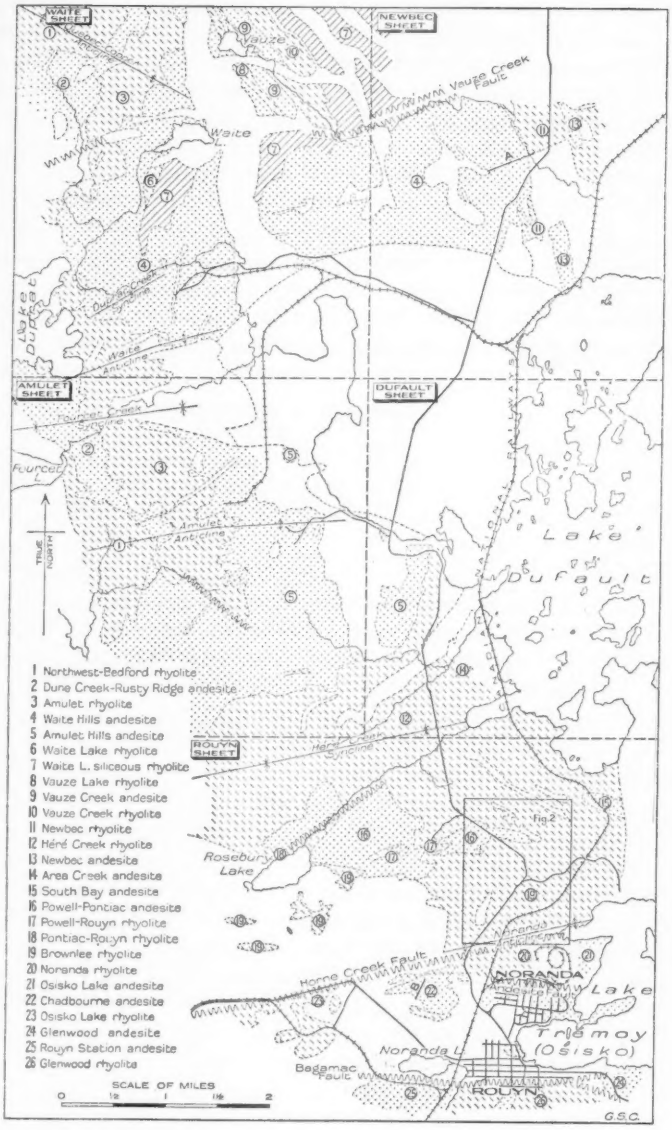


FIGURE 1.—*Lava flow belts and major structural features of Noranda district*

to include in the Keewatin most or all of the volcanics; to the Timiskaming have been assigned most of the sediments and in some localities subordinate amounts of interstratified lavas. It has also been assumed that the Timiskaming overlies the Keewatin and is separated from it by an important unconformity. The work of recent years has shown, how-

TABLE 1.—*Succession of formations*

| | |
|---|---|
| Cenozoic | |
| Postglacial | Stratified clay and sand |
| Glacial | Boulders, gravel, sand, and boulder clay. |
| Proterozoic | |
| (Late pre-Cambrian) | |
| Keweenawan? | Diabase and gabbro dikes. |
| Archean | |
| (Early pre-Cambrian) | Syenite porphyry dikes and masses |
| | Albite granite (alaskite) |
| | Granodiorite |
| | Quartz diorite, diorite, gabbro, diorite porphyry, albite granite, grandiorite. |
| Abitibi series (Keewatin) | Siliceous rhyolite |
| | Rhyolite, rhyolite flow breccia, rhyolite tuff and pyroclastic breccia. |
| | Andesite, andesite flow breccia, andesite tuff and pyroclastic breccia. |
| | Chert. |

ever, that this classification is to a considerable degree lithological and that most of the sedimentary rocks of the Archean are not so definitely separated from the volcanic members. All the surficial rocks in the area of the Noranda special maps are volcanic and hence belong to the Keewatin as usually defined.

In the region east of Noranda a synclinal belt of graywacke with some conglomerate which Gunning and Ambrose have named the Cadillac group rests in apparent conformity on the volcanics. Farther to the northeast another belt of graywacke with subordinate beds of conglomerate, the Kewagama group according to Gunning and Ambrose, is interstratified conformably with the volcanics. South of Noranda a wide belt of schistose graywacke and conglomerate, which the writer some years ago named the Pontiac series, extends from near the Ontario boundary eastward for over 100 miles. The writer has recently discovered that the southern part of this belt composed of schistose graywacke, although considerably folded, has its top to the north. It is overlain on the north unconformably by conglomerate-graywacke which in turn, except for minor folds, has its top to the north. The conglomerate-graywacke is separated from the volcanics and the Cadillac group to the north by a zone of intense faulting. Gunning and Ambrose have called this entire

southern belt Kewagama on the assumption that it is the northern (Kewagama) belt of sediments reappearing in the southern limb of a syncline, but the fault zone and the unconformity within the belt make this correlation uncertain.

The Keewatin volcanics are intruded by various igneous rocks in the form of dikes, sills, stocks, and batholiths. The most extensive intrusives from oldest to youngest are (1) quartz diorite, gabbro, and related rocks; (2) granodiorite; (3) albite granite; (4) syenite porphyry; and (5) diabase and gabbro. All these except the last are Archean. In addition to the more important intrusives (Table 1) there are numerous dikes, sills, and small masses of a great variety of other rocks. Many of these are cut by dikes of diorite, whereas others intrude the diorite. Dikes of andesite and rhyolite intrude the albite granite, and one lamprophyre was observed to intersect the granodiorite.

## ABITIBI SERIES

### LITHOLOGICAL CHARACTER

*General statement.*—The Keewatin volcanics, for which the local name Abitibi series is proposed, belong lithologically to three divisions: (1) massive (albite) rhyolite, (albite) rhyolite flow breccia, rhyolite tuff and rhyolite pyroclastic breccia; (2) siliceous (albite) rhyolite; and (3) andesite, andesite flow breccia, andesite tuff, and pyroclastic andesite breccia. The siliceous rhyolite (Fig. 1, no. 7) is a unique rock originally part of a single belt now broken into three areas, partly by faulting and partly by the intrusion of diorite. The number of belts of the rhyolitic rocks of group 1 and the andesitic rocks of group 2 is about the same, but the total thickness of the rhyolite is somewhat greater than that of the andesite.

*Rhyolite.*—The rhyolitic lava belts consist almost wholly of highly acidic rocks having the composition and physical characteristics of typical rhyolite. They, however, vary considerably in appearance on the weathered surface and locally are more basic than normal rhyolite. The predominant phrases are massive and pale gray-weathering; mottled, hackly, or mesh-weathering; bluish gray-weathering; or spherulitic. Most of these on freshly broken surfaces are fine-grained, gray to dark gray, and flintlike. They fracture conchoidally. The massive rhyolite commonly contains phenocrysts of both quartz and feldspar. Intimately associated with the massive phase are less uniform varieties. One variation is a cavernous-weathering type that originally was probably vesicular. Another phase which has a hackly weathered surface is brecciated where freshly broken. A third common variation from the normal massive

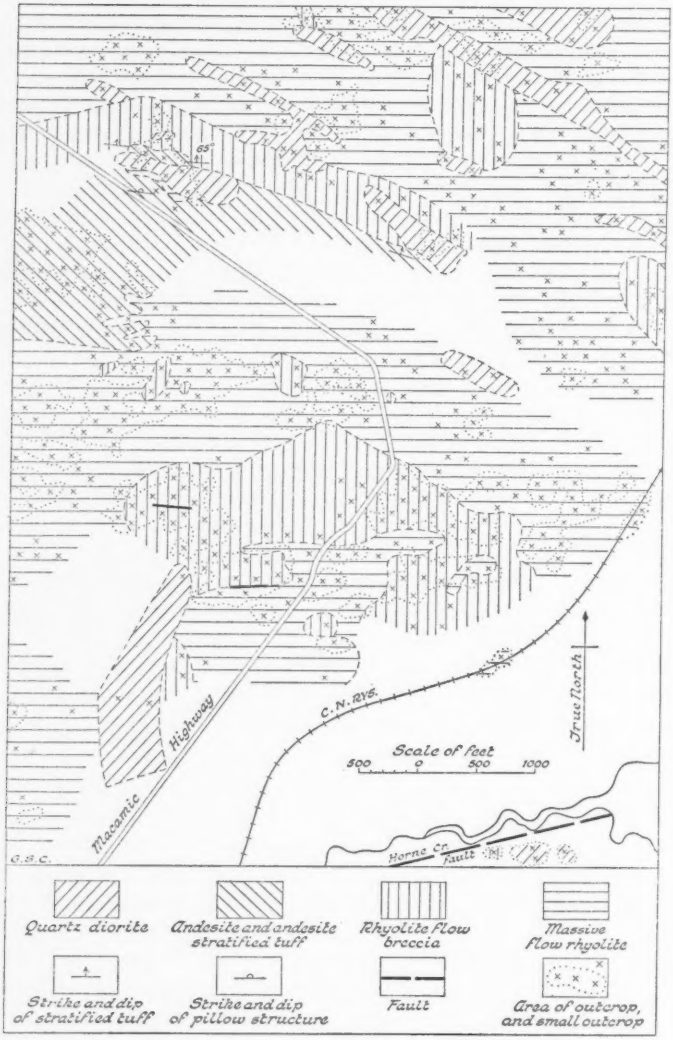


FIGURE 2.—*Plan of area north of Noranda*

Showing irregular distribution of rhyolite flow breccia in massive rhyolite. Location of area indicated in Figure 1.

rhyolite is a dull gray-weathering phase which is intersected by minute fractures giving the rock a meshlike appearance. In the belts of rhyolite are local areas of considerable extent in which weathered surfaces reveal round spherulites $\frac{1}{16}$ inch or less in diameter. These spherulitic masses pass transitionally into normal rhyolite and are probably local variations within the rhyolite.

In all the rhyolite belts there are irregular areas composed of angular fragments of rhyolite enclosed in a rhyolitic matrix. The presence of amygdules, porphyritic texture, and flow lamination within both the matrix and the fragments, and the similarity in the composition of the fragments and the matrix prove that this rock is not a pyroclastic breccia but a flow breccia. Where this breccia is separated from the other phases of the rhyolite on a map, it is found that it for the most part is scattered at random in irregular areas without control by the structure of the lava belt to which it belongs (Fig. 2). In some rhyolite flow breccias many of the fragments are sharply defined on one side and indefinite on the other (Pl. 2, fig. 2). This suggests that the sharply defined margin is the upper, and the indefinite margin the lower side of crust formed on the lava. Flow lamination is present in both the massive and brecciated phases but not so extensively as in the siliceous rhyolite.

Under the microscope the rhyolite and rhyolite flow breccia are seen to consist chiefly of phenocrysts of quartz and albite, enclosed in a matrix of granular quartz and albite. In many thin sections spherulitic, radial, micropegmatitic intergrowths of quartz and albite are also present within the matrix. Some of the quartz phenocrysts are irregularly enlarged with quartz or by a zone of micropegmatite. Most of the rhyolite has undergone varying degrees of alteration to sericite in disseminated microlites or aggregates of microlites, and green chlorite in scattered aggregates. Where alteration has been more intense, aggregates of carbonate are common. In some of the rhyolitic belts there are local dark greenish-gray andesitic-looking phases in which pillow structure is usually present. Unlike the normal rhyolite this consists of rodlike crystals of plagioclase having the optical properties of oligoclase. All these variations grade into normal rhyolite or rhyolite flow breccia and are apparently a part of the same lava flow as the associated rhyolitic rocks.

In places, chiefly in the southern part of the Rouyn map area, belts of fragmental rhyolitic material are stratified in part and are therefore presumably pyroclastic. The most extensive areas of pyroclastic breccia and tuff lie in the western part of the wedge-shaped fault block which includes the Noranda ore deposits. These consist of zones of stratified tuff or fine breccia up to 10 feet wide interbedded with unstratified breccia at intervals of 10 to 40 feet. One zone crosses the fault block in a south-

easterly direction, whereas the other lies farther west along the north side of the Andesite fault which forms the south boundary of the fault block. The western zone originally was probably continuous with that to the east but has been bent back to the west by an easterly thrust along the Andesite fault. The Noranda ore deposits lie in the broken apex of this fold.

*Siliceous rhyolite.*—This rhyolite differs from the rhyolite of the belts previously described in having a much higher silica content and in the entire absence of the flow breccia so abundant in the more typical rhyolite belts. For the most part it is uniform, pale gray to greenish-gray weathering, but in places has a hackly appearance or is broken by numerous irregular rusty depressions which suggest that the rock was originally vesicular or scoriaceous. In much of the siliceous rhyolite lamination is strikingly developed. The laminae range in width from $\frac{1}{20}$ to $\frac{1}{100}$ inch and lie $\frac{1}{20}$ inch or less apart. In places they occur throughout an entire exposure in curving zones up to 10 feet wide and continuous for 20 or even 30 feet. In many places spherulitic structure was also noted. Microscopically the siliceous rhyolite consists of numerous fine micropegmatitic intergrowths of quartz and feldspar, in part spherulitic, intermingled with irregular grains of granular quartz. In some thin sections phenocrysts of quartz and albite, for the most part less than a millimeter in diameter, are present. All the quartz is minutely cracked.

*Andesite and andesite flow breccia.*—The belts of volcanics belonging to this group consist almost wholly of gray-green to rusty-brown weathering typical andesite of uniform composition but variable appearance. The most common types are massive andesite, pillow andesite, rounded andesite breccia, and angular to subangular andesite breccia. The massive andesite is uniform-textured and occupies one-third or more of the total area mapped as andesite. In some localities it forms an entire lava flow or lies along either the top or bottom of a flow, but more commonly it is found in irregular masses of varying size within pillow andesite. In the region southeast of the town of Rouyn the interiors of some flows of andesite are coarse and dioritic looking (Fig. 3C). Under the microscope the andesite consists chiefly of elongated rodlike crystals of oligoclase to basic andesine enclosed in a matrix of abundant fibrous amphibole, magnetite, and in some sections a few grains of granular quartz. Chlorite and epidote are also commonly present as alteration products.

Andesite tuff and pyroclastic breccia occur in only a few restricted areas. The most extensive is the stratified tuff along the north margin of the Powell-Pontiac andesite belt (Fig. 1, no. 16). This zone consists

FIGURE 1. ROUNDED ANDESITE FLOW BRECCIA
East end of Waite Hills, Newbec area

FIGURE 2. RHYOLITE FLOW BRECCIA
"Ribbon type" in Vauze Creek rhyolite belt, Waite area.

FLOW BRECCIA IN ANDESITE AND RHYOLITE

FIGURE 1. IN ANDESITE OF RUSTY RIDGE BELT, AMULET AREA

FIGURE 2. IN RHYOLITE OF HÉRÉ CREEK BELT
Adjacent to Flavrian Lake road, Rouyn area.

COLUMNAR JOINTING IN ANDESITE AND RHYOLITE

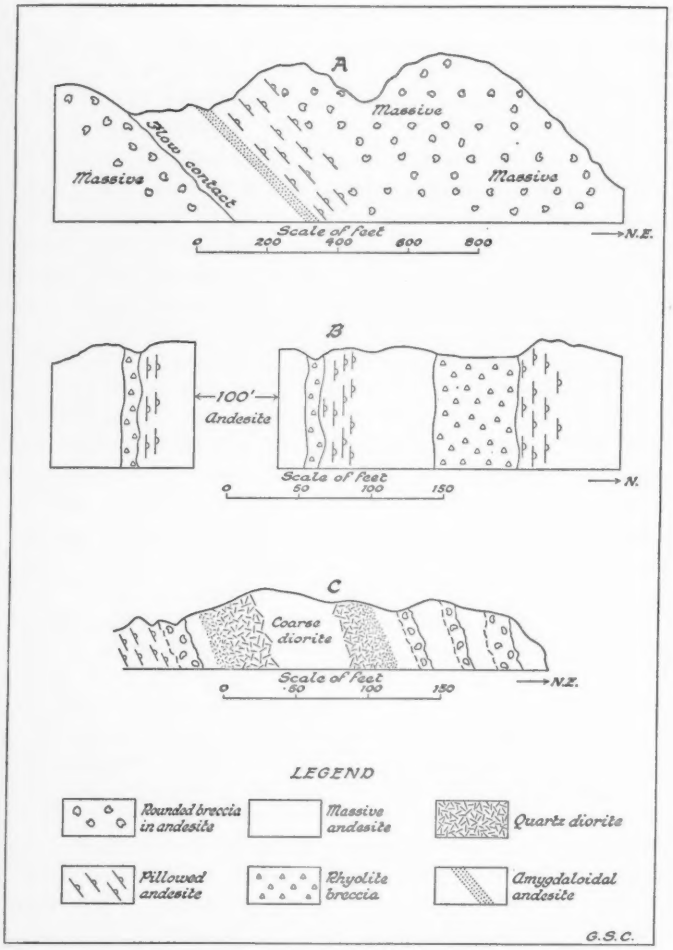


FIGURE 3.—*Cross sections through Keewatin (Abitibi) lavas*

A—andesite and andesite flow breccia near east end of Waite Hills, Newbec area
B—Chadbourne andesite and rhyolite breccia south of Tailings Lake, Rouyn area
C—andesite and andesite flow breccia, adjacent to Val d'Or road, South Rouyn village.
Location of Sections A and B indicated in Figure 1.

of light greenish-gray beds $\frac{1}{16}$ to $\frac{1}{8}$ inch wide alternating with darker beds $\frac{1}{8}$ inch to 2 inches wide. The zone is 100 feet wide and outcrops at intervals for over 1000 feet.

In numerous localities within the andesite and in one place on its contact with rhyolite (Pl. 5, fig. 1) there are zones of banded and laminated chert. These are usually not more than 3 feet thick. The maximum thickness observed was 15 feet. The chert obviously lies on flow contacts and was presumably deposited from thermal springs.

### STRUCTURAL FEATURES

*General statement.*—The features of the Abitibi volcanics used in determining their structure were (1) the bedding of the pyroclastic members; (2) columnar jointing; (3) pillow structure; (4) flow contacts within the lava belts; (5) the lava belts; and (6) cross-bedding in stratified tuff. In places in the southern part of the area the strike and dip were determined from the stratification of the interbedded tuff and pyroclastic breccia, but the breccia occurs in few localities and affords no information as to whether the dip is normal or overturned. Columnar jointing with uniform inclination occurs over extensive areas in some of the andesite and the rhyolite belts (Pl. 3). It is probable that the columns were originally vertical, and hence their deviation from the vertical indicates the dip of the lava flow.

*Pillow structure.*—The most extensively developed structures in the volcanics and the most useful in determining their flexures are the forms known as pillows. They are present in over half the andesite exposed within the area mapped. They do not occur in typical rhyolite having a granular or micropegmatitic quartz-albite matrix but were observed in five localities in the basic phase of the rhyolite. They are most common at or near the bottom of the lava flows (Fig. 3B) but may occur at the top or throughout an entire flow. In some flows a zone of massive andesite 20 feet or more thick occurs between the pillowed zone and the base of the flow. In the unusual flow shown in Figure 3A, the basal massive zone is over 160 feet thick.

These pillowlike forms are so well known that they need not be described in detail, but some of their more characteristic features in the Noranda area shall be noted. The margins of the pillows in many places are noticeably finer-grained than their interiors; in many pillows a zone of amygdules lies along the margin or less commonly along the top. Where amygdules are present throughout the entire pillow, they are more abundant along its margin. In many places, zones of lamination parallel the borders of the pillows. In some parts of the area mapped, the pillows have a buff or deep green zone up to 3 inches wide along their margins. These in some places are marked by intersecting grooves on their weathered surface similar to that observed along the borders of some of the basic dikes or sills.

Throughout the Noranda map area wherever pillow structure is seen in vertical section most of the pillows have round tops and are either flattened on their under side to a bunlike form or have adjusted themselves to the V-shaped depressions between the tops of underlying pillows, the form resembling that of a balloon. In the rock face shown in Figure 1 of Plate 4, there are three types of pillows: (1) large mattresslike pillows (in the lower right-hand corner of the photograph) having flat surfaces both above and below, (2) bun-shaped pillows, and (3) balloon-shaped pillows. One pillow of the "balloon" type rests on two bun-shaped pillows and has therefore a modified form with two V-shaped extensions at its base. At least two-thirds of the pillows in Figure 1 of Plate 4 are of the "bun" type, and between one-third and one-quarter of the "balloon" type. The bun and balloon forms of the pillows are seen fully only in sections across the lava flow. In sections parallel to the dip the pillows are round or nearly round. The flattening of the pillows on their under side to form "bun structure" and the adjustment of the pillows to the tops of underlying pillows were noted by the writer in the district northeast of Noranda in 1911. It was pointed out at that time that from the form of pillows it is possible to determine not only the strike and dip of the lava but also the upper side of the flow. Thus, in the photograph of Chadbourne andesite (Pl. 4, fig. 2), it is obvious that the lava flow has been folded into a vertical position, and, inasmuch as the photograph was taken looking east, the top of the flow is to the north. The forms of pillows are now being used extensively to determine the tops of lava flows in the Canadian pre-Cambrian. It was an invaluable aid in working out the structure of the volcanics in the Noranda area.

Pillow structure has been observed in lavas interbedded with sediments in numerous localities throughout the world. Furthermore, Tempest Anderson describes the formation of pillows in lava where it flowed into the sea from Matavanu, one of the volcanoes in the Samoan islands. For these reasons most geologists have concluded that the structure forms where lavas of basic to intermediate composition come in contact with water. Others believe that it forms subaerially, and some have even reported its occurrence in dikes.

Much of the disagreement regarding the origin of pillows, as Stearns, Stark, and McKinstry have recently pointed out, arises from confusion in the definition of the structure. Many who have concluded that pillows form subaerially assume that the pahoehoe of the Hawaiian lavas, and similar forms, are pillows, but pahoehoe is less uniform, much less discrete, and much flatter. Stearns states after 8 years of observation of Hawaiian lavas that pahoehoe surfaces are not pillows. In the

Keewatin lavas of the Noranda district the pillows are discrete masses formed in succession in an upward direction from the bottom of the flow, each succeeding pillow having adjusted its form to the pillow or pillows directly beneath it. It is scarcely conceivable that such forms could develop in magma confined in a dike.

Throughout the southern part of the Canadian Shield pillow structure is common in Keewatin lavas. The absence of interbedded limestone or other well-sorted typically marine sediments suggests that the lavas may have poured out on land. On the other hand, the uniform stratification in the pyroclastic material and other clastic sediments interstratified with the lavas indicates that standing bodies of water, presumably lakes, were common on their surface. The extrusion of the lavas terrestrially, therefore, does not prove, as some have assumed, that the lavas were not extruded into water. In regions where the climate is pluvial lava flows would almost certainly obstruct the rivers to form numerous lakes. In the Noranda district the presence of interbedded pyroclastic material indicates that the lavas were extruded into water, but the evidence for the subaqueous extrusion of the pillowed lavas of the Keewatin generally is not so conclusive as in the case of lavas interstratified with sediments known to be of marine origin.

*Flow contacts within the lava belts.*—Flow contacts are both numerous and conspicuous in the andesite of the Abitibi series. They are less common or indiscernible in the rhyolite. The most typical features that distinguish the flow contacts are: (1) a zone of breccia along the top of the flow, ranging from 3 inches to 20 feet in thickness (Pl. 2, fig. 1; Pl. 5, fig. 2); (2) laminated chert along the contact (Pl. 5, fig. 1); and (3) a zone of lamination up to 4 inches wide along the base of the flow above the contact or below the contact in places where breccia is absent at the top of the flow. Individual flow contacts can be followed in successive outcrops for 2000 feet or more. Beyond this distance they either are cut off by intrusives or faults, are hidden by drift, or terminate. In detail the line of contact is usually definite but irregular or sinuous. In general the contacts strike north-south in the northern part of the area and east-west in the southern. The tops of the flows, as shown by the presence of the zone of breccia beneath the contact, are to the eastward in the northern part of the area and to the north in the southern part. The dip ranges from 20 to 45 degrees in the northern part and is vertical or nearly vertical in the southern.

*Lava belts.*—Owing to the weathered, bare condition of the rocks in the Noranda area, there is little difficulty in distinguishing the successive

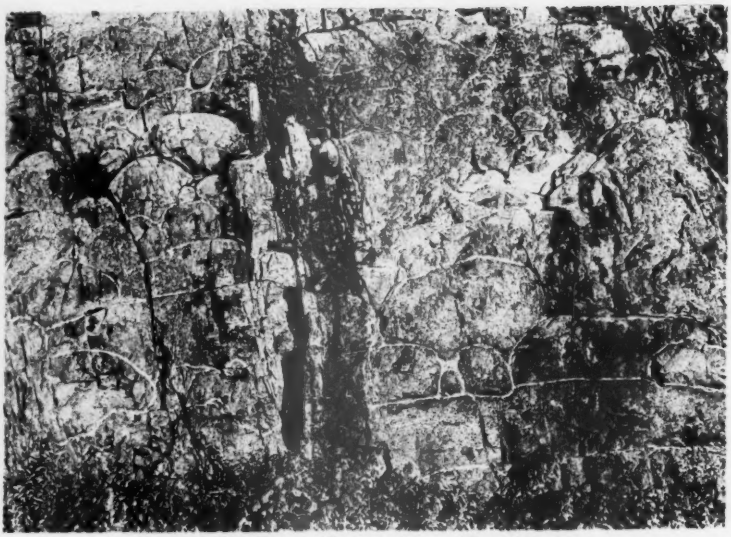

FIGURE 1. LOW-DIPPING FLOW OF AMULET HILLS ANDESITE
Southwest of No. 4 shaft, Amulet mine, Amulet area.

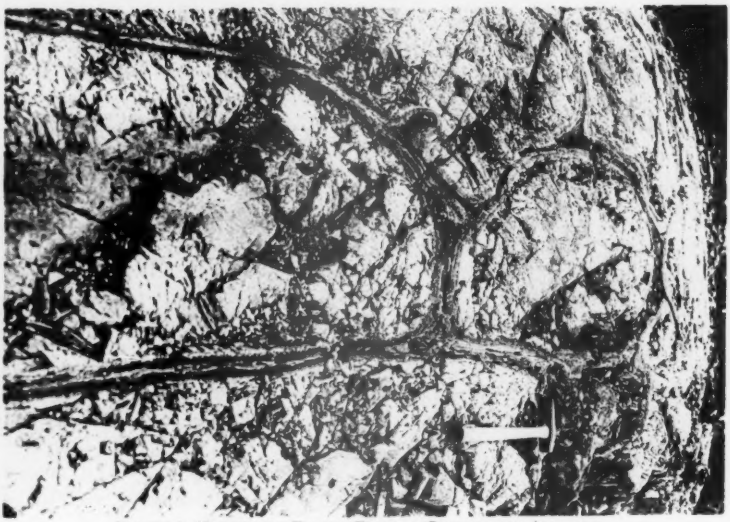

FIGURE 2. VERTICALLY FOLDED FLOW OF CHADBOURNE ANDESITE
Chadbourne property west of Noranda, Rouyn area.

PILLOW STRUCTURE IN ANDESITE

FIGURE 1. STRATIFIED CHERT
Resting on Rusty Ridge andesite and overlain by Amulet rhyolite west of Amulet mine, Amulet area.

FIGURE 2. FLOW CONTACT
Andesite (right) resting on rounded andesite breccia flow (top) adjacent to Val d'Or road, South Rouyn village.

CHERT, ANDESITE, AND ANDESITE FLOW BRECCIA

lava belts, the rhyolite being in general pale gray in contrast to the gray-green weathering andesite. Because the surface of lavas is usually irregular it is not possible to determine the strike and dip of the lava belts definitely from the outcrops of their contacts alone. From the attitude of pillows, from the flow contacts within the lava belts, and from diamond drill holes bored through the contacts in places, it is known, however, that in the northern part of the area mapped they have been folded with moderate intensity into eastward-plunging anticlines and synclines, whereas in the southern part they have been closely compressed into vertical or nearly vertical folds plunging more steeply toward the east. The folds include four great anticlines and three intervening major synclines, the names of which are shown in Figure 1.

The separation of the lava belts on the maps afforded evidence of numerous faults, the presence of most of which would otherwise be unknown. The most important of these are the Vauze Creek, the Horne Creek, and the Andesite faults. The zone of stratified tuff and pyroclastic breccia in which most of the ore of the Noranda (Horne) mine occurs is cut off on the north by the Horne Creek fault, but a similar zone recurs on the north side about 2 miles to the east. This fact, together with the similarity of the South Bay andesite belt on the north side of the fault to the Chadbourne andesite on the south (Fig. 1), indicates that the rocks on the north side of the fault have moved eastward with respect to those on the south. Similarly a comparison of the rocks on the opposite sides of the Vauze Creek fault shows that the rocks on the north have moved eastward at least 4000 feet with respect to those on the south.

The manner in which intrusions of diorite follow the faults and cut across the folds is proof that much of the folding and faulting occurred before the diorite was intruded. On the other hand the presence of shear zones in the diorite along the faults indicates that faulting recurred after the intrusion of the diorite. In the Rouyn map area a large mass of albite granite is cut off sharply along its southern margin by the Horne Creek fault. Except for two small areas directly opposite the mass to the north, there is no granite anywhere on the south side of the fault. From these relationships it is concluded that faulting recurred after the granite was intruded and that the south side moved vertically downward with respect to the north. Farther eastward along the fault, north-south dikes of later diabase in the Powell and Noranda properties are displaced 700 feet in opposite directions along the Horne Creek fault. This displacement shows that faulting also recurred after the diabase was intruded, and as in the case of the albite granite mass the movement was about vertical.

TABLE 2.—*Thickness of volcanic rocks*

| Name of belt | Assumed average angle of dip | Thickness (in feet) | | |
|---|---|---|---|---|
| | | Minimum | Maximum | Average |
| Northwest-Bedford rhyolite | 30° | | | 1600 |
| Dune Creek-Rusty Ridge andesite | 30° | 000 | 1400 | 700 |
| Amulet rhyolite | 30° | 600 | 2500 | 1550 |
| Waite Hills-Amulet Hills-Vauze Creek-Powell-Pontiac andesite | 30° (northern part) | 800 | 11,000 | 5900 |
| Vauze Lake rhyolite | 30° | 150 | 400 | 275 |
| Vauze Creek-Waite Lake rhyolite | 30° | 000 | 750 | 375 |
| Waite Lake siliceous rhyolite | 30° | 000 | 2000 | 1000 |
| Andesite on northeast part Waite area | 30° | 080 | 120 | 100 |
| Newbec-Here Creek rhyolite | 30° | 2000 | 4000 | 3000 |
| Newbec andesite | 30° | 250 | 400 | 325 |
| Area Creek andesite | 45° | 200 | 200 | 200 |
| South Bay andesite | 90° | 000 | 8000 | 400 |
| Brownlee rhyolite | 60° | | | 4000 |
| Total | | | | 19,425 |
| **SOUTH OF HORNE CREEK FAULT** | | | | |
| Noranda rhyolite | 90° | | | 3000 |
| Osisko Lake andesite | 90° | | | 1000 |
| Chadbourne andesite | 90° | 2000 | 2000 | 2000 |
| Osisko Lake rhyolite | 90° | 2000 | 2000 | (2000) |
| Glenwood andesite | 90° | | | |
| Rouyn Station andesite | 90° | | | |
| Glenwood rhyolite | 90° | | | |
| Total | | | | 6000 |

SUMMARY

| | Rhyolite | Siliceous rhyolite | Andesite | Total (feet) |
|---|---|---|---|---|
| North of Horne Creek fault | 10,800 | 1000 | 7625 | 19,425 |
| South of Horne Creek fault | 3000 | | 3000 | 6000 |
| Total | 13,800 | 1000 | 10,625 | 25,425 |

THICKNESS

The lava belts are so irregular and have been so much faulted that it is difficult to determine their thickness accurately. An estimate of the thickness of each flow belt is included in Table 2. In making these estimates it has been assumed that the several areas of siliceous rhyolite outcropping in the Waite map area were originally part of a single belt

and that the Powell-Pontiac andesite in the north-central part of the Rouyn area is the southern continuation of the Amulet Hills andesite. Some of the lava belts north of the Horne Creek fault are almost certainly repeated south of the fault. For this reason only a part of the belts south of the fault is included in the estimate of thickness of lavas in the area.

## CONCLUSIONS

In the Noranda district, Quebec, an area of 45 square miles recently investigated by the writer is of considerable importance because it includes the copper-gold-bearing sulphide ore deposits of the Noranda (Horne), Amulet, and Waite-Ackerman-Montgomery mines, as well as the gold-bearing quartz veins of the Powell Rouyn property. The total value of the mineral production from the district in 1940 amounted to $25,717,000.

The rocks of the area, except for some dikes of diabase and gabbro, all belong to the Archean and consist for the most part of alternating belts of Keewatin rhyolite and andesite, intruded in places by dikes or masses of diorite, granodiorite, or albite granite. Although the lavas have been highly folded, their original character is well preserved so that it is possible from the trend and attitude of the lava belts, from flow contacts within the lava belts, and, most of all, by the determination of tops from the form of pillows and the zones of breccia along the tops of flows to ascertain their structure.

From these data it was determined, as shown in Figure 1, that the lavas of the area have been folded in four major eastward-plunging anticlines, with intervening synclines, and have been displaced by numerous easterly- or northeasterly-trending faults along most of which the rocks on the north side have moved east with respect to those on the south. It has also been determined that the movement along some of these faults was recurrent extending from early Archean to very late pre-Cambrian time. The earliest movement on the Horne Creek fault was a thrust, the north side moving eastward about 2 miles with respect to the south. The later movements were normal, the south side moving downward vertically or nearly vertically with respect to the north.

The ore deposits of the Noranda (Horne) mine lie within a westward-pointing wedge-shaped fault block delimited on the north by the vertical or nearly vertical Horne Creek fault and on the south by the Andesite fault. The latter converges westward toward the Horne Creek fault and dips 82° N. Zones of stratified tuff and pyroclastic breccia extend northwesterly across the fault block and along the north side of the Andesite fault. The relationships suggest that the tuff and breccia form an east-

ward-plunging anticline the south limb of which has been dragged westward along the Andesite fault. Numerous southeasterly-dipping zones of shearing extend northeasterly into the fault block from the Andesite fault. The ore deposits occur partly in broken rhyolite but most extensively in the tuff and pyroclastic breccia where these zones of deformation intersect the anticline.

Within the area examined are at least nine belts of rhyolitic and eight of andesitic volcanics having a total thickness of over 25,000 feet; about two-fifths are andesite, and the remainder rhyolite. The presence of original breccia or lamination at the tops of the flows shows that the lavas were extruded without intervening intervals of erosion. The Keewatin volcanic rocks of the area are, therefore, part of a single series for which the local name Abitibi series is proposed. The upper and lower limits of this series lie outside the area mapped and are as yet unknown.

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# GEOLOGIC HISTORY AND CORRELATION OF THE JURASSIC OF SOUTHWESTERN OREGON AND CALIFORNIA

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## ABSTRACT

The generally accepted correlations of the Jurassic of southwestern Oregon and California are as follows: Galice and Mariposa, Dothan and Franciscan, Myrtle and

Shasta. The Dothan is said to be younger than the Galice. On the other hand it has been stated that the Dillard, the lower part of the original Myrtle, is the equivalent of the Franciscan. Since the Dillard unconformably overlies Dothan and Galice, it is clear that one or the other of these correlations must be erroneous. Evidence is presented to show that the Dothan is older than the Galice and cannot be the same age as the Franciscan. The Dillard, as originally defined, includes Knoxville (Upper Jurassic) and possibly infolded Lower Cretaceous, but the great bulk of the Dillard is to be correlated with the Franciscan. The Myrtle, as mapped and defined, includes beds of both Upper Jurassic and Lower Cretaceous age and at least one important unconformity. The name should be redefined and restricted or abandoned.

The Nevadan orogeny took place in the late Kimmeridgian and uplifted the Sierra Nevada and depressed the belt now occupied by the Coast Ranges. In the geosyncline thus formed the Franciscan and Knoxville were deposited. There is no unconformity between the Franciscan and Knoxville; these groups were deposited during the Portlandian and Tithonian.

At the base of the Lower Cretaceous is a widespread unconformity. The diastrophism which closed the Jurassic is called the Diablan orogeny.

## INTRODUCTION

### GENERAL STATEMENT

This paper attempts to clarify confusion caused by many conflicting published statements regarding the age, succession, and correlation of certain rocks in southwestern Oregon and California. It also presents evidence as to the age of the Franciscan group and the dating of certain important events in Pacific Coast history. It is in no way a criticism of the pioneer workers in this region who under great difficulties, on foot and horseback, contributed so greatly to our knowledge of many remote and inaccessible regions but is an attempt to reconcile the results of several investigators working in different areas and often only slightly familiar with the geologic features of other regions. Needless confusion has been introduced by later writers with little first-hand knowledge of the areas discussed.

The writer has spent a number of years on a field study of the Franciscan and Knoxville of the Coast Ranges of California and the metamorphosed Jurassic rocks of the Sierra Nevada and is familiar with the existing literature. Practically all of his field work has been in California, but 10 days were spent in an examination of several type sections in southern Oregon. The various correlations that have been made are, to say the least, confusing, and a reader unfamiliar with the problem would gain an impression that the entire matter is hopelessly jumbled. Although present knowledge does not justify a statement of the precise age of several of the units, the author believes the correlations and successions proposed are essentially correct. They are based on first hand information of the areas discussed and on an intimate acquaintance with all the important literature. Only brief descriptions of the various units will be given since they are fully described in papers

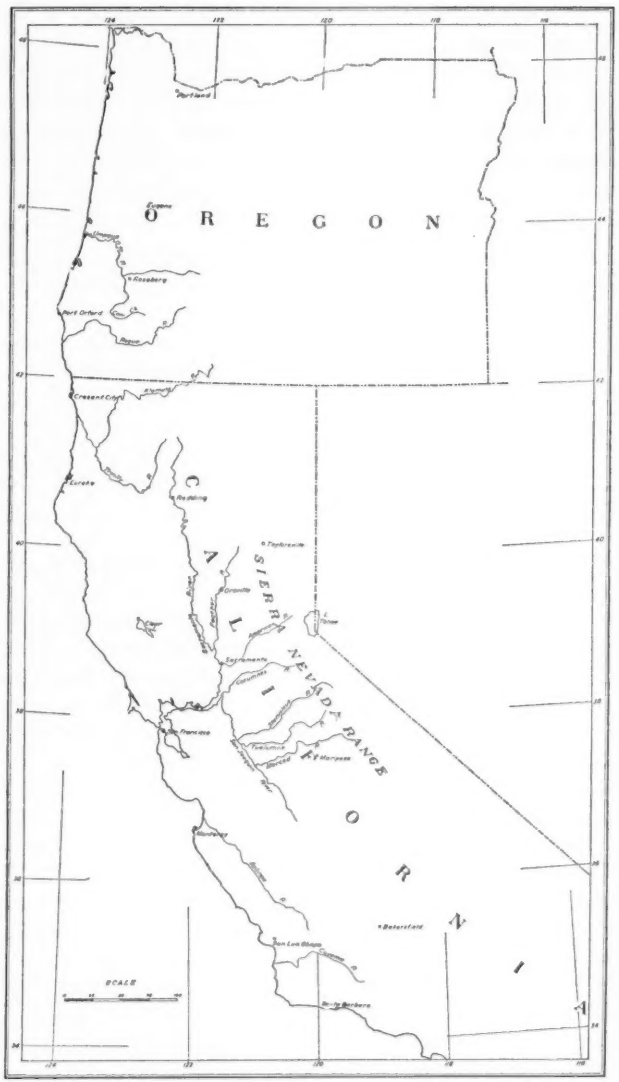


FIGURE 1.—*Map of California and Oregon*
Showing the location of features mentioned in the text.

now in preparation on the Franciscan and Knoxville and on the Jurassic of the Sierra Nevada.

The writer is indebted to Professor Siemon W. Muller and Dr. F. M. Anderson for the determination of fossils. Grateful acknowledgment is made to the Board of Research of the University of California for the partial support of this and related problems.

## STATEMENT OF THE PROBLEM AND UNITS DISCUSSED

The writer makes no attempt to discuss the whole Jurassic of California; only those units in California that appear to have correlatives in southwestern Oregon are considered although mention is made of the older Jurassic.

Several Jurassic formations have been described in the southwestern corner of Oregon in Curry, Josephine, Jackson, Douglas, and Coos counties, and several diametrically opposed correlations of the Oregon and California units have been made. In southwestern Oregon the two unquestioned Jurassic formations are the Dothan and Galice; the Myrtle is stated to be Lower Cretaceous but as mapped and described it includes at least one important unconformity and sediments and igneous rocks of Upper Jurassic age. The Dillard series was separated from the Myrtle (Louderback, 1905), but the name has not been generally used or accepted.

In the Coast Ranges of California the Franciscan and Knoxville (restricted) are Upper Jurassic and in the western part of the Sierra Nevada the Monte de Oro beds and the Amador and Mariposa groups are Jurassic. The Jurassic rocks of the Coast Ranges and those of the Sierra Nevada are never in contact in California, although they approach each other closely in the Klamath Mountains; only in southwestern Oregon do they come into contact.

The Galice of Oregon has been correlated with the Mariposa of the Sierra Nevada of California, while the Dothan and Dillard (lower part of the Myrtle) have been correlated with the Franciscan. They cannot both be equivalent to the Franciscan since the Dillard lies unconformably on and overlaps Dothan and Galice and is much less metamorphosed. The Dothan is said to be younger than the Galice but in the Sierra Nevada the Mariposa (equivalent to the Galice) is younger than the Amador and Monte de Oro beds which appear to be equivalent to the Dothan. This paper attempts to clarify these involved and mutually contradictory correlations and to fix the age and relations of the Franciscan; if this purpose can be attained certain important diastrophic events and their effects can be better understood and more accurately dated.

## MYRTLE, DILLARD, AND FRANCISCAN

The Mesozoic of southwestern Oregon has been studied by Diller (1903; 1907; 1908a), Diller and Kay (1924), Knowlton (1910), Louderback (1905), Fontaine (1905), Stanton (1895), and Ward (1900); Crickmay (1931) and Goranson (1924) have drawn certain broad conclusions without first hand knowledge of the region. The present ideas regarding the age and succession of the beds may be found under the various groups and formations in a work by Wilmarth (1938). The following section indicates the generally accepted succession and correlation adopted by the United States Geological Survey:

| | Southwestern Oregon | California |
|---|---|---|
| Lower Cretaceous | Myrtle | Horsetown<br>Knoxville |
| Upper Jurassic | Dothan<br>Galice | Franciscan<br>Mariposa |

The writer does not accept the above succession for southwestern Oregon.

Diller (1907, 1908a) correlated the Galice of Oregon with the Mariposa slates of the Sierra Nevada of California on the basis of lithology and fauna and the Dothan with the Franciscan of the Coast Ranges of California, this latter correlation being without any basis save the presence of cherts in both and the supposed position of the Dothan above the Galice. The Myrtle was considered to be Lower Cretaceous, equivalent to the Knoxville and Horsetown in California. Louderback (1905) pointed out the heterogeneous elements included in Diller's Myrtle and considered the lower part, containing clastic sediments, radiolarian cherts, glaucophane and related schists, and basic and ultrabasic intrusives, to be the equivalent of the Franciscan, naming it the Dillard series. Diller (1907; 1908a) revisited the area and reported "Lower Cretaceous" fossils at a number of localities, considering this sufficient evidence for retaining the Dillard in the Lower Cretaceous Myrtle; judging by the literature this appears to be the generally accepted opinion at present. As "Lower Cretaceous" was used at that time this conclusion was justified, but as a result of work by Knowlton (1910), Smith (1909), Pavlov (1907), Haug (1898), and Anderson (1932; 1933a; 1933b) the lower Knoxville part of the original Myrtle must be placed in the Upper Jurassic. From a study of Diller's map (1907) and a visit to the region the fossils all appear to come from the upper part of the "Dillard" and

the lower part of the Myrtle. The fossils reported by Diller are *Aucella piochii* and *A. crassicollis;* therefore, according to Anderson, (1932; 1933b; 1938) both the Upper Jurassic and the Lower Cretaceous (Paskenta stage) are represented. *Aucella piochii* is now generally regarded as being confined to the Upper Jurassic, and this agrees with the writer's field evidence. The form usually called *Aucella "crassicollis"* is, according to H. G. Schenck, really *Aucella crassa,* which is stated to be confined to the Lower Cretaceous (Paskenta stage) although the writer does not consider this to be firmly established. While *Aucella "crassicollis"* is nearly always found in beds above those containing *Aucella piochii* the two occasionally occur in the same beds, below the unconformity at the base of the Lower Cretaceous. Both commonly occur together in the basal conglomerate of the Paskenta stage of the Lower Cretaceous, but in this case *Aucella piochii* occurs in blocks and boulders of reworked Knoxville (Upper Jurassic). The writer is not convinced that *Aucella "crassicollis"* is a marker for the Paskenta or is confined to the Lower Cretaceous. It would be unsafe to state that the beds in the vicinity of Dillard, Oregon, were Lower Cretaceous solely on the basis of *Aucella "crassicollis."* Lower Cretaceous beds may be present in small synclines in the area covered by Louderback's Dillard group, but unquestionably the great bulk of the Dillard series and that part of Diller's Myrtle in this locality are lithologically identical with the Franciscan of the Coast Ranges of California. The discovery of fossils, nearly all species of *Aucella* formerly regarded as indicative of the Lower Cretaceous, is not sufficient justification for including the beds in the vicinity of Dillard, Oregon, Louderback's Dillard series) in the Myrtle and regarding them as Lower Cretaceous.

The Myrtle formation, as mapped and defined, is a heterogeneous assemblage ranging from Upper Jurassic to late Lower Cretaceous and including at least one important unconformity. In the Port Orford quadrangle, Oregon, both Dothan and Galice, as well as Franciscan, Knoxville, Paskenta, and Horsetown are included in the Myrtle; in the Roseburg quadrangle, Oregon, the Myrtle, as mapped, includes Franciscan, Knoxville, and Horsetown; in the Coos Bay quadrangle, Oregon, the Myrtle appears to be largely, if not entirely, Franciscan. The inclusion of so many heterogeneous elements in the Myrtle is not surprising when the nature of the exposures, the intricate folding and faulting, the nature of the base maps, and the rapidity of the mapping are considered. In the field the separation of these units over wide areas could only be accomplished by slow and detailed mapping on much better base maps than now exist. Nevertheless, the Myrtle formation as mapped and as accepted by the United States Geological Survey (Wilmarth, 1938)

is a heterogeneous assemblage containing both Upper Jurassic and Lower Cretaceous sediments and at least one unconformity. In the literature the Myrtle is Lower Cretaceous and is stated to be the equivalent of the Shasta of California but as shown on various published maps it includes a wide variety of older rocks. Therefore, the name Myrtle either should be redefined or abandoned. If redefined it should include only Lower Cretaceous beds and not the Upper Jurassic Dillard which is actually the equivalent of the Franciscan of California. There is a marked unconformity between the Upper Jurassic and Lower Cretaceous both in California and Oregon, and debris of the Upper Jurassic is abundant in the Lower Cretaceous conglomerates.

## GALICE AND MARIPOSA

### GENERAL CONSIDERATIONS

Unquestionably the Galice, at least at the type section on Galice Creek and northward, is equivalent to a part if not all of the Mariposa of the Sierra Nevada of California. They are identical lithologically, have been metamorphosed to the same degree, and contain the same fauna. Both are characterized by *Aucella erringtoni* and similar species of *Ammonites*. Acid and intermediate volcanics, tuffs, flows and agglomerates, are interbedded with them. In places volcanics make up a larger part of the section than slates and sheared sandstones; the volcanics were submarine, coming from many local submarine centers. A broad belt of similar slates and volcanics containing *Aucella erringtoni* extends southward from the type section of the Galice into northern California as far south as the Trinity River where unmetamorphosed Franciscan occurs within half a mile of the slates. Of all of the correlations of various parts of the Upper Jurassic of southwestern Oregon with the California section this is the most certain; Mariposa and Galice are unquestionable correlatives.

### AGE OF GALICE AND MARIPOSA

Hyatt (1892), Smith (1894; 1909; 1910), Turner (1894; 1896), Reeside (1919), and Crickmay (1933) have discussed the age of the Mariposa. Since the discovery of fossils in these slates, sandstones, and tuffs in the latter part of the last century the Upper Jurassic age of the Mariposa has been generally accepted but not all have agreed as to the precise stage. Fossils are fairly abundant at many widely spaced localities in the Mariposa, but often they are crushed making specific determination difficult. The best and most complete collections were obtained in Calaveras County on Texas Charley Gulch, a tributary of the Stanislaus River, and southeastward at Reynolds Ferry and Bostick Bar on the Stanislaus. Species of *Aucella* and *Ammonites* are fairly abundant, and

*Belemnites* is occasionally found. The writer and 35 students spent two days collecting in this area, but no new species were added to the list of those discovered nearly 50 years earlier. Although somewhat distorted, much of the material from this region is determinable. Among the species of *Ammonites* from this locality is one described by Hyatt (1892), Smith (1909), and Reeside (1919). Smith called it *Cardioceras whitneyi*, Hyatt *Cardioceras dubium*, and Reeside *Amoeboceras dubium*. Muller (personal communication) agrees with Reeside that it is *Amoeboceras* rather than *Cardioceras* and hence is upper Oxfordian. This is associated in the same beds with *Aucella erringtoni* (and related species of *Aucella*) and with *Perisphinctes virgulatiformis* Hyatt, which Crickmay (1933) called *Virgatosphinctoides virgulatiformis* and placed in the very late Kimmeridgian. Since the preservation of the *Amoeboceras* described and figured by Reeside and the same form in Muller's collection is far better than the *V. virgulatiformis* described and figured by Crickmay (*cf.*, Reeside, 1919, Pl. XXIV, figs. 24, 25 with Crickmay, 1933, Pl. 16, figs. 24, 25) the writer considers that upper Oxfordian is more reasonable than late Kimmeridgian; this is also in far better agreement with the important diastrophic events which occurred after the deposition of the Mariposa and before the deposition of still later Jurassic sediments. Crickmay (1933) described and figured Hyatt's *Aucella erringtoni* var. *arcuata* from Moffat Bridge on the Tuolumne River as *"Buchia arcuata"* and dated the beds containing it as "Late Upper Jurassic, Tithonian" (1933, p. 54-55). Regarding the supposed Tithonian age of this species Crickmay makes the following statement:

> "This is immensely interesting and important, because formerly no Mesozoic rocks younger than Kimmeridgian have been known to occur in the Sierra Nevada. The slabs of rock containing *'Aucella' arcuata* are composed of green sandstone and shale, very different from the slate of the Mariposa formation, in which most of the Upper Jurassic fossils of the Sierra Nevada are found. Evidently these slabs came from beds which have been included in the Mariposa formation of many authors but which, in the writer's opinion, would be better separated as a younger formation. This unit has, of course, remained unnamed and, indeed, hitherto unrecognized as a distinct unit. A host of problems arise at once: Have these beds of latest Jurassic age correlatives in California or in the regions immediately beyond? What is their relationship to the great batholiths and to the high-angled structure of the Sierra Nevada? These questions are of fundamental importance. They bear directly upon the late Jurassic history, not only of California but of the whole Cordillera. Their solution will answer in some measure the outstanding problem: Did the Nevadan orogeny take place at one time? If so, when? Or did it come as several successive steps? Again, if so when? Finally, how was the deformation related to the great intrusions? These are all field problems and must be turned over to the field geologist."

Following this closing suggestion the writer, who already had spent a great deal of time in the revision of the Jurassic stratigraphy of the Sierra Nevada, revisited the Moffat Bridge locality. Three days were spent in restudying the section, and three days, with five students, in

collecting at this locality. Not only were collections made at the surface but shallow pits were dug and a large collection, chiefly species of *Aucella*, was obtained.

Before discussing the fossils the writer wishes to take up the matter of the "green sandstone and shale," a description based on material collected over 50 years earlier and unchecked by field observations. Anyone familiar with the Mariposa, or for that matter any similar type of black slate subjected to weathering in a region such as the Sierra Nevada, knows that the black slates and sandstones weather to a greenish-gray color; below the surface or on fresh exposures the greenish color changes to the usual black. This phenomenon was described by Whitney (1879, p. 45) as occurring in the Mariposa in general and in the belt which runs through Moffat Bridge in particular. At the Moffat Bridge locality, situated near the lower contact of the Mariposa, are tuffs, flows, and agglomerates into which the sediments grade and which the writer has named the Logtown Ridge member of the Amador group. The slates and sandstones at this locality are lithologically identical with the Mariposa throughout the Sierra Nevada and are in the same belt which extends to the northwest to and beyond the Texas Charley Gulch and Stanislaus River area. To the southeast it extends to and beyond the fossiliferous horizon at Hell Hollow on the Merced River which yielded a fauna similar to the one found on Texas Charley Gulch. However, the fossils at Moffat Bridge are not nearly so well preserved as at the other localities.

The fossils were submitted to S. W. Muller who reported as follows (personal communication, 1936):

> "Now a few words about the material which you were kind enough to send me for study. I must admit at the outset that the preservation of the *Ammonites* precludes even a guess as to the exact age of the rocks beyond the fact that they are probably Upper Jurassic. As to the *Aucellas*, I am willing to guess and in this connection I wish to say that your material from near Moffat Bridge contains almost exclusively the radially striated forms which according to some authorities (Sokolov, 1908) are classed together in the *A. bronni* group (*including A. erringtoni*) and are regarded as characteristic of Upper Oxfordian and Kimmeridgian. There is a suggestion that in collection the radially striated forms are associated with a few smooth forms which makes it more likely that the rocks in question are Kimmeridgian rather than Oxfordian. However, if these rocks are the same as those from the Texas Charley Gulch locality, I would be more inclined to regard them as Oxfordian. I have just looked over some of the old material in our collections from the 'Texas' ranch which shows affinities with the European Oxfordian. In either case, it is clear, that I cannot agree with Crickmay that the rocks at the Moffat Bridge locality are Tithonian—provided we are talking about the same rocks."

There can be no question as to the locality, for the original description clearly states that the fossils were obtained on the south bank of the Tuolumne River at the southeast end of Moffat Bridge. Although the bridge has long since washed away both abutments remain. The

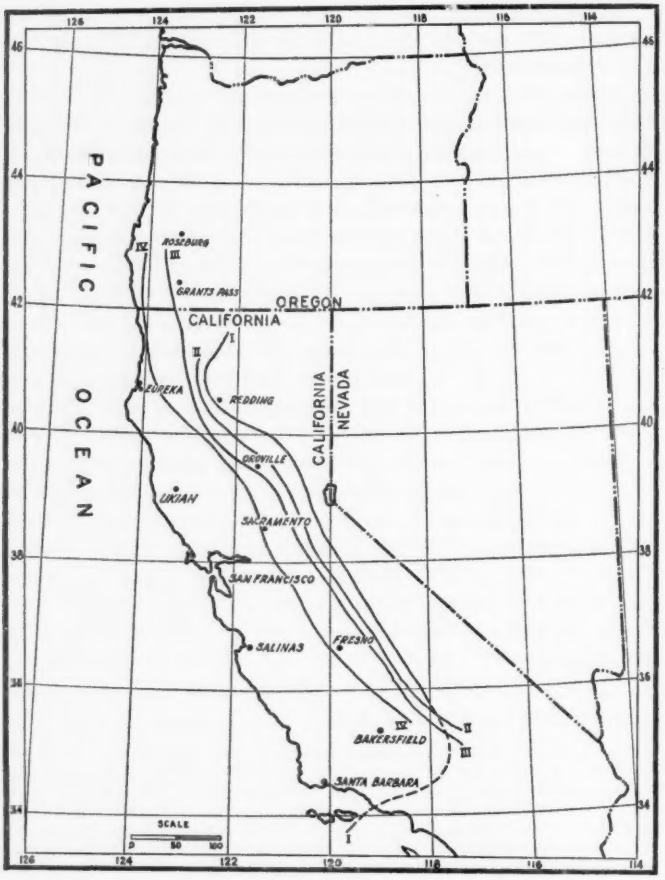


FIGURE 2.—*Map of California and Oregon*

Showing the approximate western limits of I, Upper Triassic sea; II, Lower Jurassic sea; III, Middle Jurassic sea; IV, Kimmeridgian sea.

writer's collections come from a stratigraphic thickness of about 50 feet on both banks of the river.

The writer has examined his large collections from the Texas Charley Gulch and Bostick Bar localities and has found occasional smooth Aucellas from these localities where they are associated with Amoeboceras and abundant radially striated Aucellas. In Hyatt's original description of *Aucella erringtoni* var. *arcuata* he states that this form was found

not only at Moffat Bridge but also on the Stanislaus River. The writer considers that the evidence favors an upper Oxfordian age for the Mariposa at Moffat Bridge. The Mariposa as a whole may be upper Oxfordian and lower Kimmeridgian; there is no definite evidence that the Mariposa extends into the upper Kimmeridgian. Since the Moffat Bridge beds are evidently true Mariposa and not a younger formation of Tithonian age most of Crickmay's questions are answered. The Mariposa along the Tuolumne River is intruded by serpentine within half a mile of the fossil locality and the serpentine is known to be older than the Sierran granodiorite; the Mariposa is, of course, commonly intruded by the granodiorite. The question raised by Crickmay regarding the nature and time of the Nevadan orogeny will be discussed in a later section of this paper. The Galice is undoubtedly the same age as the Mariposa since both contain the same species of *Ammonites* and *Aucella*.

## RELATION BETWEEN GALICE AND DOTHAN

Diller (1907) named and described both the Galice and the Dothan concluding that the Dothan was younger although the Dothan lies spatially below the volcanics which underlie and are interbedded with the Galice; the entire section, Galice, volcanics, and Dothan, was thought to be overturned. This relationship was reaffirmed by Diller and Kay in 1924 and, judging by the literature, the Galice is generally accepted as being older than the Dothan.

This supposed relationship, and the correlation of the Dothan with the Franciscan, were so opposed to the very definite relations and succession observed in California, as well as to paleobotanical evidence, that the writer spent 10 days in an examination of the Oregon type sections. He studied the section from Reuben Station to West Fork Station (Dothan Post Office) on the Southern Pacific Railroad along Cow Creek, including the type section of the Dothan and the locality near Reuben at which the largest collection of Galice fossils was obtained; the type section of the Galice, on Galice Creek and the Rogue River, and the Graves Creek section, as well as the Dillard area, were visited.

Between Dothan Post Office (West Fork Station) and Reuben, Cow Creek cuts approximately across the strike of the beds and the section is well and practically continuously exposed in natural stream cuts and in artificial cuts along the railroad. Diller described the Dothan as consisting of shales and sandstones and the Galice as made up largely of slates, implying a marked difference in the stage of metamorphism. Actually, however, the Dothan and Galice have been equally deformed; the original shales in each have been converted into slates and the sandstones somewhat sheared. The metamorphism, however, has not been

sufficiently severe to obliterate the original bedding, especially where there are rapid alterations of fine and coarse clastics, and no difficulty was encountered in determining the tops and bottoms of beds by the usual methods—channeling, graded bedding, cross-bedding, and ripple marks. Dothan sediments are well exposed for 4½ miles southeast of Dothan Post Office. The general strike of the bedding varies from north-south to N. 30° E.; the beds have been folded and the dip is variable but averages approximately 50° SE. Immediately southeast of Dothan the small folds are open and only slightly asymmetric; proceeding to the southeast the small folds become steeper and are strongly asymmetric or overturned toward the northwest. The beds are cut by numerous small thrust faults and shear planes which dip southeasterly. In this section the Dothan consists of black slates, sheared sandstones, and occasional lenses of impure black limestone and thin conglomerates. The sediments are usually thin-bedded but there are a number of sandstone layers up to 100 feet thick; the sandstones, slightly predominant over the slates, are decidedly arkosic. Under the microscope they are seen to be composed chiefly of angular grains of quartz, plagiocase, and orthoclase, the total feldspar being fully as abundant as the quartz; occasional grains of zircon and sphene and shreds of mica, chlorite, and epidote are present. Rock fragments are abundant, especially in the coarser layers, and consist of quartzite, schist, slate, aplite, and fine-grained volcanics. The sediments were derived from granodiorites and various crystalline rocks and were deposited in shallow water in a subsiding trough.

Aside from the small drag folds, the top of the beds is always toward the east or east-southeast, as shown by the previously mentioned details of bedding as well as by the position of the drag folds; the beds are only locally overturned on the short northwestern limbs of the small anticlines. Undeniable evidence proves that the progression is from older to younger beds going toward the east or east-southeast—the direction of the dip—and that the Dothan sediments in the type section are not overturned except locally on drag folds.

The first volcanics appear 4½ miles southeast of Dothan Post Office (1000 feet west of mile post 517). Bombs of vesicular basalt and tuffaceous sediments are interbedded with the ordinary black Dothan slates. The main body of flows, tuffs, and agglomerates appears about 600 feet east of mile post 517. Occasional thin layers of black slates and dark arkosic sandstones are interbedded with the volcanics, and 2 miles southeast of the first appearance of the volcanics black slates, 200 feet thick, are interbedded with schistose tuffs and flows. The volcanics are largely schistose rhyolite and andesite flows, tuffs, and agglomerates; they are

folded and occasionally dip toward the northwest. At Reuben Station fossiliferous Galice slates appear interbedded with the same volcanics; beneath the bridge over Cow Creek at Reuben Station, and above the fossiliferous slates, are good exposures of sheared rhyolite agglomerates.

The unmistakable relations in this section indicate that the Dothan grades upward into the volcanics, which contain interbedded sediments; fossiliferous Galice slates are interbedded with the volcanics still higher in the section. The Dothan is older than the Galice and, at the type section at least, is separated from it by volcanics which grade downward into the former and upward into the latter. The sediments interbedded with the volcanics could be arbitrarily placed either with the Dothan or the Galice. As in California the volcanics interbedded with the sediments occur as great lenses at various horizons from the lower part of the Dothan to the top of the Galice. No thick section is completely free from volcanics but the ratio of volcanics to sediments varies greatly; in some places volcanics predominate and replace thick sections of sediments and in others the dynamically metamorphosed sediments predominate. Rhyolitic and andesitic volcanics predominate, but basalts are not uncommon, especially in the Dothan. The sections on Cow and Graves creeks and on the Rogue River are very similar to many sections on the rivers of the Sierra Nevada of California, west of the Mother Lode. These will be briefly described in a later section.

## FRANCISCAN AND KNOXVILLE

### GENERAL CONSIDERATIONS

Before discussing the various suggested correlations a brief statement will be made regarding the nature, age, and relationship of the Franciscan and Knoxville even though this subject has been previously discussed by the writer (1941) and a definite statement made regarding the very controversial matter of the age of the Franciscan.

The Franciscan consists of a heterogeneous assemblage of rock types—clastic, chemical, organic, pneumatolytic metamorphics, volcanics, and intrusives. The bulk of the Franciscan, especially the lower part, consists chiefly of arkosic sandstones with relatively thin shale partings; the upper part is made up of the same type of sandstone, a greater proportion of dark shales, pillow basalts, basalt and andesite agglomerates and tuffs, radiolarian cherts and occasional red and gray foraminiferal and algal limestones. Fossils are very rare, aside from the poorly preserved and recrystallized Foraminifera in the limestones, and recognition of the Franciscan is made on the basis of lithologic types. Significantly, where fossils have been found the beds have, as a rule, been called Knoxville. The writer has visited all the localities from which fossils have

been reported in the Franciscan; these will be discussed in a future paper on the Franciscan and Knoxville. In the northern and central Coast Ranges the Franciscan is usually overlain by dark shales and sandstones which have been called Knoxville. As used here Knoxville refers only to the lower (Upper Jurassic) part of the original Knoxville, which included both Upper Jurassic and Lower Cretaceous sediments. As will be shown later there is a widespread break between the Upper Jurassic Knoxville and the Lower Cretaceous.

The Knoxville is usually described as consisting of dark shales and sandstones and an occasional conglomerate. This is an adequate description of the bulk of the formation, but it ignores certain important and widespread types in the lower part. In northern California, in Lake County, north of Wilbur Springs in Colusa County, and in several places in the central Coast Ranges, where the relations are not obscured by faulting the Franciscan visibly grades into fossiliferous Knoxville sediments. In the Wilbur Springs region typical Franciscan cherts, shales, and pillow basalts grade upward into volcanic tuffs and agglomerates which are interbedded with dark shales and an occasional flow of pillow basalt. Typical Knoxville fossils, species of *Belemnites* and *Aucella,* have been found in the dark shales and sandstones above, below, and interbedded with the pillow basalts and volcanic agglomerates. In southern San Luis Obispo County an abundant "Knoxville" fauna consisting of species of *Ammonites, Aucella, Belemnites,* and brachiopods interbedded with and below pillow basalts and radiolarian cherts. These beds are unconformably overlain by a conglomerate carrying a large Lower Cretaceous fauna. The Knoxville fauna in this locality is stated to be equivalent to a very high horizon in the Knoxville of northern California.

Ordinarily, a distinction may be made between Franciscan and Knoxville, and the two may be mapped as distinct units. However, where the contact is exposed it is gradational and if traced for any distance would transgress time since in some places volcanism continued almost to the end of the Jurassic. In San Luis Obispo County, volcanism, and the deposition of radiolarian cherts, continued to a very late stage in the Jurassic, resulting in Franciscan lithologic types, but in other regions where volcanism ceased at an earlier time sandstones and shales, having a distinct Knoxville lithology and fauna, were deposited. This complexity makes it difficult to discuss the Franciscan and Knoxville as separate units except where lithologic differences are marked.

The Franciscan and Knoxville are intruded by great ultrabasic and basic sills, laccoliths, dikes, and plugs. The ultrabasic peridotites and dunites were largely converted to serpentine during and immediately after their intrusion, but the cores of some of the larger plugs are still unaltered.

These intrusions occasionally caused local pneumatolytic metamorphism and the development of glaucophane and related schists about their margins.

Many conflicting statements have been made regarding the relation of the Franciscan to the basement complex of the central Coast Ranges of California and the relation of this complex to the plutonics and metamorphics of the Sierra Nevada. Because of the many severe diastrophisms suffered by the central Coast Ranges all of the contacts between the Franciscan and the Sur series and Santa Lucia plutonics are faults. However, the conglomerates in the Franciscan, especially in the western part of the central Coast Ranges, contain abundant debris of the Gabilan marble, Sur schists, and Santa Lucia plutonics derived from the land mass to the west. Furthermore, the Jurassic of the Sierra Nevada also contains abundant debris of these rocks. The schists and plutonics of the central Coast Ranges are older than those of the Sierra Nevada; they are either early Paleozoic or pre-Paleozoic. These rocks were deformed, dynamothermally metamorphosed, and deeply eroded before the deposition of the Amador and Mariposa of the Sierra Nevada which are definitely older than the Franciscan.

### AGE OF FRANCISCAN AND KNOXVILLE

An abundant fauna occurs in the Knoxville, and little doubt exists as to the age of the sandstones and shales lying above the Franciscan. Because the relations between the Franciscan and Knoxville have never been recognized previously, the age of the Franciscan has been uncertain. Usually it is considered Upper Jurassic but it has been suggested that it might be as old as the Triassic or even the Permian; it even has been suggested that it might occupy an unnamed interval between the Jurassic and Cretaceous. The many unjustifiable correlations (Reed, 1933) of the Franciscan with other and older beds in the Sierra Nevada and southern California have introduced a great deal of unnecessary confusion.

The Franciscan and Knoxville are confined to the Coast Ranges of California and are never in contact with the Mariposa of the Sierra Nevada. However, in the Coast Ranges (Klamath Mountains) of northern California a belt of slates and schistose volcanics lithologically identical with the Mariposa and the Galice contains *Aucella erringtoni* which is common to both formations. A broad belt of unmetamorphosed Franciscan sandstones, shales, basalts, and cherts lying within half a mile of and for a long distance parallel to the regionally metamorphosed equivalent of the Mariposa and Galice leaves little doubt that the unmetamorphosed Franciscan is later. Furthermore, in Oregon, unmetamorphosed Franciscan lies unconformably on both Dothan and Galice.

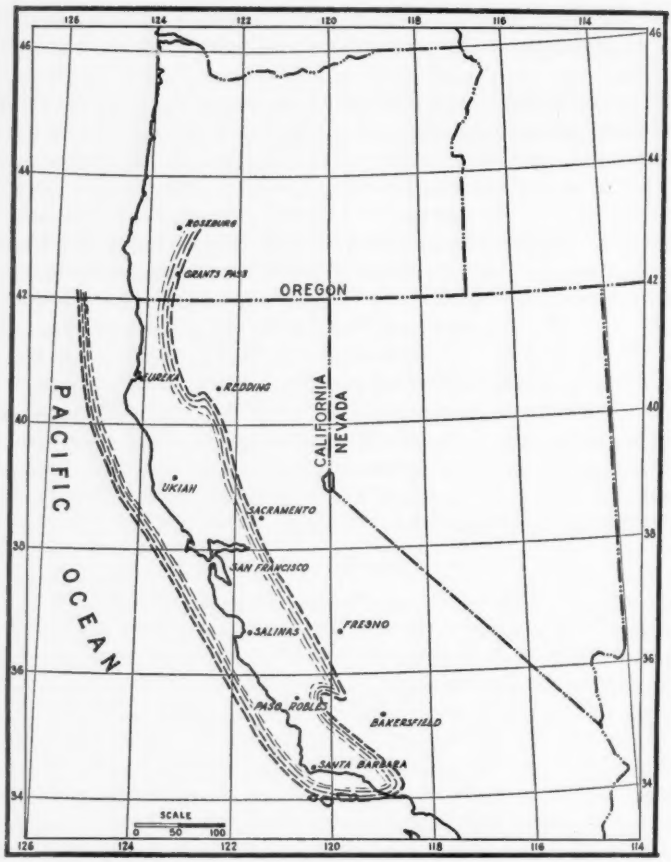


FIGURE 3.—*Map of California and Oregon*

Showing the approximate position of the late Upper Jurassic sea in which the Franciscan and Knoxville were deposited.

Since the Mariposa is known not to extend above the Kimmeridgian (probably not above the lower Kimmeridgian) and the abundant Knoxville fauna is stated by Anderson (1932; 1933a; 1933b) to be upper Portlandian and Tithonian (Aquilonian)[1] the Franciscan, is post-lower

[1] The use of Portlandian and Tithonian together, although sanctioned by custom, is, perhaps, unfortunate since it is a combination of British and Continental terms. There is not complete agreement on the nomenclature of this part of the Upper Jurassic. By some the Tithonian is used as a stage name for everything in the Jurassic above the Kimmeridgian and is divided into Bononian and Aquilonian. These latter are essentially the same as Portlandian and Tithonian as used here.

Kimmeridgian and pre-Aquilonian. However, as shown previously, certain lithologic phases, usually regarded as being typical of the Franciscan extend upward almost to the top of the Jurassic. The writer believes he has collected sufficient physical evidence, notwithstanding the usual absence of diagnostic fossils, to establish the age of the Franciscan within comparatively narrow limits. The Franciscan is not a catch-all of uncertain age, to be correlated with various wholly unrelated beds in the Sierra Nevada and southern California, but occupies a definite position in the geologic column and has a definite geographic distribution whose limits are fairly well known. It was deposited in a geosyncline developed between a newly risen or rejuvenated land mass west of the present coast and the ancestral Sierra Nevada. The high and rugged western land mass, made up of plutonic and metamorphic rocks, was subjected to mechanical attack under rigorous climatic conditions and contributed great floods of coarse arkosic sands to the basin; as this western land mass was worn down, chemical weathering became more important, and a greater proportion of fine clastics were contributed. The shales in the upper part of the Franciscan and in the Knoxville represent detritus from the worn down western land mass. Erosion of the unmetamorphosed surface beds of the Sierra Nevada may have contributed debris to the Franciscan-Knoxville geosyncline. Widespread submarine volcanism did not begin until the geosyncline had sunk several thousand feet and a considerable thickness of shallow water sediments had accumulated. The maximum development of radiolarian cherts coincided with the maximum volcanism. The cherts were deposited in shallow water as shown by the fact they are frequently interbedded with coarse arkosic sandstones and even conglomerates. Volcanism resulted in submarine flows of pillow basalts and in basaltic and andesitic agglomerates and tuffs. In general, volcanism ceased before the deposition of the Knoxville shales, but in some localities pillow basalts and agglomerates are found above fossiliferous Knoxville shales, and in one locality (southern San Luis Obispo County) volcanism apparently continued to the close of Knoxville time.

## SUPPOSED EQUIVALENCE OF DOTHAN AND FRANCISCAN

The various proposed correlations are very confusing. Diller (1907; 1908a) correlated the Jurassic Dothan with the Franciscan, and Louderback (1905) correlated the lower part of Diller's Lower Cretaceous Myrtle with the Franciscan. Since these conflicting correlations have been made by geologists thoroughly familiar with northern California and southwestern Oregon they must be carefully analysed.

The correlation of the Dothan with the Franciscan, as proposed by Diller (1907; 1908a), Diller and Kay (1924), and as accepted by the Geological Survey (1938), is not supported by the nature of the two and their stratigraphic position. In the first place the lithologic successions of the two are quite different and many types characteristic of the Franciscan are lacking in the Dothan. Secondly the Dothan is everywhere in a more advanced stage of metamorphism than the great bulk of the Franciscan. The shales of the Dothan have been converted by regional metamorphism into slates identical with those of the Galice and Mariposa. The glaucophane and related schists of the Franciscan and Knoxville are the result of local pneumatolytic contact action on the borders of ultrabasic and basic intrusives. They occur in very small areas and grade within a few feet into unaltered sediments; the schistosity is a result of the original bedding and not an effect of pressure. These schists resulted from local, not regional, metamorphism. In the third place the Dothan underlies the Galice and both unconformably underlie the lower part of Diller's Myrtle, the Dillard series of Louderback (1905). Furthermore, unmetamorphosed Franciscan occurs within half a mile of dynamically metamorphosed Mariposa-Galice slates west of the Trinity River in northern California. The writer can see no justification whatever for the correlation of the Dothan with the Franciscan.

The correlation of the lower part of the Myrtle, the Dillard series, with the Franciscan rests on a much firmer foundation. (1) The Dillard contains all the usual lithologic types characteristic of the Franciscan, coarse and fine clastics, rhythmically bedded radiolarian cherts, glaucophane schists, pillow basalts, red and gray foraminiferal limestones, and basic and ultrabasic intrusives. (2) The general succession is the same, the lower part of both being largely coarse-to-medium-grained graywackes, followed by volcanics, graywackes, radiolarian cherts, black shales, and thin limestones. This succession is characteristic of the Franciscan from Santa Barbara County on the south to southwestern Oregon on the north. (3) The Franciscan may be traced, with but one minor interruption due to faulting, from the type section about San Francisco Bay into the Port Orford quadrangle in Oregon. The writer has collected *Aucella piochii* from impure limestone lenses in black shales intruded by gabbro and serpentine in the Dillard region. While this fossil is usually regarded as characteristic of the lower Knoxville it has been found in California in both the lower Knoxville and the upper part of the Franciscan, as these are usually defined on the basis of lithologic types. Probably the Dillard series, as originally defined, includes sediments that would ordinarily be called Knoxville, and infolded Lower Cretaceous beds may be present. However, the major part of the Dillard

series of Louderback, the lower part of Diller's original Myrtle, is undoubtedly equivalent to the Franciscan of the California Coast Ranges. These rocks show no evidence of regional metamorphism and overlap a broad belt of Dothan and Galice slates and volcanics. The writer considers that there is positive evidence that a part of the Dillard and not the Dothan is the equivalent of at least a part if not all of the Franciscan.

## AMADOR GROUP

The writer has studied the stratigraphy and structure of the Sierra Nevada, west of the Mother Lode thrust and has separated from the Mariposa a thick section of Mesozoic sediments and volcanics called the Amador group using the exposures along the Cosumnes River in Amador County. This group was originally named Tuolumne (1933) but since that was preoccupied the name Amador was substituted.

Everywhere west of the Mother Lode the Mariposa is underlain by a thick section of sediments and volcanics. In the early Sierran folios these beds were mapped either as part of the Mariposa, as Calaveras, or as amphibolite, quartz amphibolite, diabase, or porphyrite. Sections have been made on most of the rivers from Mariposa River north to the American River and the structure and stratigraphic sequence determined. The Amador sediments, which are strongly folded and frequently faulted, range in thickness from less than 5000 to more than 15,000 feet. Lateral variation and great submarine volcanic centers at various horizons sometimes make it impossible to correlate the Amador formations from one part of the Sierras to the other. Except where interrupted by serpentine and granodiorite intrusions and by faulting the base of this group is exposed as far south as the latitude of Angels Camp, between the Calaveras and Stanislaus rivers, on the flanks of a great compound anticline whose crest is made up of Paleozoic sediments (Calaveras). On the Cosumnes River 1200 feet of conglomerates and sandstones forms the base of the Amador on either side of the Calaveras. Cobbles and boulders of the underlying Paleozoic rocks, as well as schists which closely resemble the older Sur schists of the Coast Ranges, are abundant in these Amador conglomerates. Both the Amador and the Mariposa were derived from the west as shown by coarsening in that direction.

The great bulk of the Amador consists of volcanics and clastic sediments, but chemical sediments such as red and green radiolarian cherts and dense unfossiliferous limestones are common. These usually occur as thin discontinuous lenses, but on the Merced River radiolarian cherts with interbedded tuffs and shales are over 1500 feet thick. These beds, which the writer has called the Hunter Valley cherts (Taliaferro, 1933), overlie about 1400 feet of pillow basalts.

The Amador sediments and volcanics usually grade upward into the Mariposa but locally between the Merced and Mariposa rivers conglomerates at the base of the Mariposa contain abundant pebbles and cobbles of the underlying Amador.

The Mother Lode represents a great Upper Jurassic thrust along which the Paleozoic was thrust westward over the Mesozoic. Mariposa beds always lie in a strongly faulted, crumpled, and overturned syncline immediately west of the Mother Lode thrust which was of sufficient magnitude to override and cut out all of the Amador, except occasionally a few hundred feet of the uppermost beds. Both the Amador and Mariposa were removed from the upthrust eastern block; Jurassic sediments again appear far to the east, near the crest of the range, in a greatly faulted syncline which has been almost obliterated in many places by batholithic invasion. West of the Mother Lode thrust the belts of Mariposa shown on the Sierran folios usually represent synclines separated by anticlines exposing the Amador rocks.

There is little direct evidence regarding the age of the Amador except that it usually conformably underlies the Mariposa. Species of *Ammonites* have been collected by S. W. Muller and the writer on the Cosumnes River 3500 to 4000 feet above the base, but they are so badly crushed and distorted that the only statement that can be made is that they are Jurassic.

The Amador of the Sierra Nevada of California occupies the same stratigraphic position as and is lithologically identical with the Dothan of southwestern Oregon. The section of the Dothan and Galice on Cow Creek is very similar to many sections in the Sierran rivers west of the Mother Lode thrust. In both regions the sediments (Dothan and Amador) grade upward into volcanics, which contain interbedded sediments and which grade upward into sediments (Galice and Mariposa). The Dothan and Amador are very unfossiliferous compared with the Galice and Mariposa. The paleobotanical evidence of the age of the Dothan will be discussed later. Although there is no direct paleontological evidence, the writer regards the Amador and the Dothan as roughly equivalent. One may extend lower in the Jurassic than the other but the upper parts of both are Upper Jurassic.

## MONTE DE ORO BEDS

A short distance northeast of Oroville, on the Feather River, black carbonaceous sandy slates, sandstones, and conglomerates, a few hundred feet thick, are preserved in the deepest part of the trough of a tightly compressed overturned syncline. These beds, traceable for only a short

distance away from the Feather River, form an isolated area more than 40 miles from any belt of Sierran Jurassic sediments. Their isolated position is not due to nondeposition of the Jurassic over the Sierran region to the east but to removal by erosion following the Nevadan revolution; the Monte de Oro beds were preserved because of their structural position.

These sediments rest unconformably on Paleozoic volcanics and have a thin basal conglomerate filled with debris of the underlying rocks. Thin sections of 20 pebbles from the basal beds were studied microscopically; 18 are andesite porphyries identical with the beds beneath, one is a quartzite, and one a thoroughly recrystallized black chert. The pebbles are well rounded to subangular and usually less than 3 inches in diameter.

The Monte de Oro beds were named and briefly described by Turner in 1896 and placed in the Jurassic on the basis of an abundant flora and a meager and poorly preserved fauna. The fauna has been studied by Stanton (quoted by Allen, 1929), Diller (1908a), and Knowlton (1910), who concluded the beds were definitely Jurassic but who was uncertain as to the stage represented. He suggested that the Monte de Oro beds might be equivalent to some of the older Jurassic sediments of the Taylorsville region. The flora was studied by Fontaine (1905) and Ward (1900) who concluded it was the equivalent of the flora of the Inferior Oölite beds of Yorkshire, England, which are Bajocian, or about Middle Jurassic. Knowlton (1910) reached the same conclusion but also correlated the Monte de Oro fauna with the Elk River, Curry County, Oregon flora, and with the floras from Big Bar and Rattlesnake Creek, Trinity County, California, all of which contain a definite Lower Cretaceous fauna. However, the three floras just mentioned are poorly preserved, considerably macerated and difficult to determine, which is in marked contrast to the Monte de Oro flora.

The sparse and poorly preserved fauna indicates little more than a Jurassic age although some evidence suggests an equivalence to the Middle Jurassic beds of the Taylorsville region. The flora, exceptionally well preserved, has been correlated with the Middle Jurassic floras of Yorkshire, England and Siberia. Hence the best available evidence indicates that these beds are Middle Jurassic and older than the Mariposa. If a Middle Jurassic age is correct, it is possible that the Monte de Oro beds are a near shore equivalent of a part of the Mormon formation of the Taylorsville region (Diller, 1908b), which is older than the Foreman formation—generally regarded as equivalent to the Mariposa. The relation of the Monte de Oro beds to the Amador group is not known since the nearest exposures of the two are 50 miles apart, but the writer

believes the Monte de Oro beds are the older, provided they are actually Middle Jurassic.

The state of preservation of the plant remains in the Monte de Oro and the character of the sediments indicate deposition very close to the western margin of the Middle Jurassic sea which extended eastward. The sediments are strongly cross-bedded and channeled and were evidently deposited in very shallow water. The sandstones are arkosic and contain an abundance of rather fresh angular feldspar. The slates originally were very sandy silts and also contain an abundance of fresh angular feldspar and quartz grains. The appearance of the plants and the lack of maceration shows that they were not transported for any distance and must have been entombed very close to the forests from which they were derived.

The Monte de Oro beds are especially important in indicating a point on the position of the western shore line of the Middle Jurassic sea. Unfortunately, these sediments have not been preserved over wide areas.

## BEDS CONTAINING THE "JURASSIC FLORA OF DOUGLAS COUNTY, OREGON"

The Jurassic flora of Douglas County, Oregon has been described by Fontaine (1905) Knowlton (1910) and Ward (1900) and all have correlated it with the floras of the Monte de Oro beds and the Middle Jurassic beds of Yorkshire and Siberia. The evidence for this correlation is presented in the publications referred to above and will not be reviewed here.

Fossil plants were obtained from 20 localities which fall into three closely spaced areas in Douglas County, Oregon: Thompson Creek on the north, vicinity of Buck Peak 1½ mile south-southeast, and Cow Creek in the vicinity of Nichols Station (long since abandoned) on the Southern Pacific Railroad 6 miles south of the Thompson Creek localities. The plants have been adequately described and figured but the descriptions of the beds in which they occur are meager and conflicting. The writer has not visited these localities but has seen large collections from Thompson Creek and Buck Peak in the Museum of Paleontology of the University of California. Thin sections of the matrix are identical with the Monte de Oro beds and the brief description given previously applies equally to both. This is also true of the preservation of the plants. Thus another point is established on the western margin of the Middle Jurassic sea, again assuming that the correlations are correct. The relation of the Jurassic plant beds to the Dothan is not definitely known; Diller (1907; 1908a) considered the beds at Nichols Station to be Dothan but included the plant beds at Buck Peak in the

"Myrtle." There is slight evidence that the plant-bearing beds lie along the crest of an anticline whose eastern flank is made up of Dothan sediments and volcanics and whose western flank is overlapped by Cretaceous and Eocene.

Sufficient evidence of the relation between the plant beds and the Dothan is lacking, but if all of the plant beds containing the "Jurassic Flora of Douglas County, Oregon" to the north and south of Buck Peak are of the same age and are actually Middle Jurassic, they appear to be older than the Dothan and bear the same relation to the Dothan as do the Monte de Oro beds to the Amador group in California. Until more positive structural, paleontological, or paleobotanical evidence is accumulated no positive statement is possible.

## JURASSIC VOLCANICS

The great thickness of volcanics is a striking feature of practically all Jurassic units of California and southwestern Oregon. The volcanics range from rhyolites to basalts, but augite andesites predominate. Practically all, if not all, are submarine as they are interbedded with marine sediments, and the water-worked tuffs occasionally contain marine fossils. They are widely distributed both in time and space; they vary greatly in thickness within the individual units, being thicker near the centers of volcanism.

Flows of basalt, pillow basalt, augite andesite, dacite, and rhyolite are numerous but pyroclastics usually predominate. All gradations occur between slightly tuffaceous sediments and pure tuffs. Volcanic breccias are practically always present; these vary from thick accumulations of angular blocks averaging more than 10 feet in diameter near volcanic centers to thin tuffaceous beds containing scattered and somewhat rounded fragments. Centers of maximum volcanism are easily recognized by the thickness of the volcanic accumulation and the coarseness of the volcanic breccias. A concrete example is the great volcanic center of the Horseshoe Bend Mountains north of the Merced River in Mariposa County. Here are volcanics, flows, tuffs, and very coarse breccias fully 12,000 feet thick. These have been followed to the southwest for 30 miles; in this distance they gradually thin and become finer grained and finally grade into well-bedded tuffaceous sediments and slates.

Occasionally the actual volcanic necks and conduits may be recognized; both breccia necks and solid pluglike bodies occur. The breccia necks occasionally carry numerous angular fragments of an older granodiorite, torn from the walls of the conduit as the volcanic material blasted its way upward. Usually the contemporaneous sediments and tuffs show great variation in strike and dip about such necks because

the bedded material was crushed against these more rigid bodies during the Nevadan orogeny.

Pillow basalts are not uncommon, especially in the Amador and Dothan and occasionally they predominate locally, as on the Merced River where successive flows of albitized pillow basalts are 1400 feet thick. Red and green radiolarian cherts are usually associated with the pillow basalts. although they may be associated with any of the volcanics except the very coarse agglomerates.

Immediately west of the Mother Lode thrust is an almost continuous belt of augite andesites and basalt flows, breccias, and tuffs in the upper part of the Amador. Eight to 10 miles to the west (formerly a much greater distance because of the shortening of the section by compression) and at the same horizon is an equally continuous belt of rhyolites and augite andesites. In the eastern belt the volcanoes erupted only augite andesites and basalts and in the western belt, near the present border of San Joaquin Valley, the submarine volcanism was chiefly rhyolitic. Until these two types were recognized as contemporaneous great difficulty was encountered in interpreting the sequence of these strongly folded and metamorphosed beds. This petrographic difference is reflected in the hypabyssal intrusives; on the east these are largely andesite porphyries and diabases and on the west quartz porphyries.

The volcanics have been folded and metamorphosed to the same extent as the sediments. The more massive volcanics, except where locally sheared and hydrothermally altered, have been little modified as to structure and texture, and original features such as amygdules, spherulites, and flow banding are readily recognized. The tuffs, especially the original vitric-crystal tuffs, whether rhyolitic, andesitic, or basaltic, have been greatly modified and converted into a variety of schists. The tuffs yielded even more readily than the original shales with which they were frequently interbedded and therefore completely schistose beds, representing the original crystal vitric tuffs, are found alternating with black slates.

## ULTRABASIC INTRUSIVES

Intrusions of peridotite and dunite, now largely serpentinized, together with their closely associated gabbroic and diabasic differentiates, are common in the Jurassic of California and southwestern Oregon. These occur as sills, dikes, plugs, and large masses of indeterminate form. The great majority of these intrusives in the Sierra Nevada are irregular sills, clearly intruded prior to the folding of the Jurassic sediments and volcanics since they are folded with them. In northern California and Oregon the large bodies of peridotite and dunite, in various stages of

serpentinization, may have been intruded after the folding. However, in the Sierra Nevada all bodies studied by the writer were intruded prior to the folding and batholithic invasion. Like the younger serpentines intrusive into the Franciscan and Knoxville of the Coast Ranges of California these ultrabasic intrusives were emplaced in a later stage of geosynclinal sinking and after the accumulation of a thick prism of sediments. In the Coast Ranges the ultrabasic rocks were evidently serpentinized immediately after, and in some cases during and even prior to their final emplacement. Knopf (1929) has shown that the serpentines of the Mother Lode region were serpentinized at a very early date, probably immediately after their emplacement.

The serpentines of the Sierra Nevada were probably emplaced during the Kimmeridgian, certainly prior to the Nevadan orogeny. The serpentines of the Coast Ranges were emplaced in the Tithonian, after the Nevadan orogeny. They are both Jurassic, but the Sierran serpentines are earlier than those in the Coast Ranges. Previous attempts to correlate these serpentines have been based on erroneous assumptions. Both were intruded in a late stage of geosynclinal sinking, but in different regions and at different times.

The serpentines of the Sierra Nevada have been slightly metamorphosed by the folding and greatly changed on the contacts of the granodiorite batholiths. Thin lenses and the margins of some of the bodies have been converted into talc schist by dynamic metamorphism.

## NEVADAN REVOLUTION

### REVIEW OF THE LITERATURE

The important and severe orogeny that folded, crushed, and thrust-faulted the Jurassic and older rocks of the Sierra Nevada has been given many names and many diverse interpretations since its importance was recognized nearly 75 years ago. Although much has been written about this diastrophism there is not complete agreement as to the exact time at which it took place or the areal limits of the area affected. Some have postulated that the entire Cordilleran region, from Mexico, or even South America, northward into Canada, or even Alaska was folded and intruded by batholiths in the mid-Mesozoic; others have limited the effects to the Sierra Nevada of California. Usually two events have been included in this orogeny, namely, the folding of the beds and the batholithic invasions,—events which may be manifestations of the same subcrustal forces but which are separated in time and in effect so that a distinction should be made between them. The time of the last severe folding is known within rather narrow limits but the plutonic invasions are less susceptible to accurate dating.

All observations show that the folding antedated the final emplacement of the great batholiths. The writer has mapped a number of folds in the Jurassic rocks of the Sierra Nevada and has traced them to the borders of stocks and off-shoots of the main batholithic area and found them again on the opposite side, sometimes as much as 8 miles away, with the same trends and the same formations exposed along them. Plutonic bodies are also found either cutting across or guided by strong Upper Jurassic thrust zones. Folding and thrusting unquestionably preceded plutonic intrusion but the length of the interval between the two events is not known. The early phases of plutonic invasion may have taken place in depth during or immediately after folding but the final emplacement of the batholiths was not accomplished until the folds and thrusts were fully developed. Plutonic intrusion may have immediately followed the folding or there may have been an appreciable or even a considerable interval between them. Broadly speaking they may be considered as component parts of a revolution, but it must be recognized that they may be separated by a considerable interval.

The ideas regarding the Sierra Nevada developed by the early Geological Survey of California were summarized by Whitney in 1879. It was recognized at that time that the youngest beds standing at high angles and showing signs of strong compressive forces were Jurassic and the oldest undeformed rocks resting on them, in the extreme western part of the Sierras adjacent to the Great Valley, were Cretaceous. He implied, but did not actually state, that the deformation took place at the close of the Jurassic. All of the Sierran folios, published near the close of the last century, stated that the disturbance which followed the deposition of the Mariposa beds was the last of the movements folding the Auriferous slate series.

Lawson (1893) first applied the name "Cordilleran" to this revolution, recognizing that folding had taken place but placing greater emphasis on the widespread plutonics. He believed that the entire Cordilleran region, from South America to Alaska and from the Wasatch Mountains to the Pacific, was affected by this "pre-Cretaceous Mesozoic revolution." Diller and Stanton, impressed by the unconformity at the base of the Cretaceous in northern California, concluded that both the Sierra Nevada and the Coast Ranges were deformed at the close of the Jurassic. In the northern end of the Sacramento Valley gently dipping Upper Cretaceous sediments rest on highly inclined Sierran bedrock on the east side and Lower Cretaceous beds lie on Paleozoic rocks on the west side. However, the same conditions do not prevail in the Coast Ranges to the south where Lower Cretaceous sediments rest on Upper Jurassic beds (Franciscan and Knoxville) that are younger than the Mariposa.

Smith (1909) believed that the Cordilleran revolution metamorphosed the Jurassic of the Sierra Nevada and the Franciscan of the Coast Ranges but took place before the deposition of the Knoxville and hence was post-Mariposa and pre-Portlandian. Blackwelder (1914) reviewed the statements concerning this orogenic period and proposed the term Nevadian, "in allusion to the Sierra Nevada," because he believed the term Cordilleran too inclusive and tended to give an erroneous picture of the diastrophism. He noted that the time of the diastrophism had been given by various writers, for areas extending from Mexico to Alaska, as ranging from Middle Jurassic to the very close of the Jurassic but believed that:

> "these differences of opinion are probably in large part due to the difficulty of correlating the faunas of such widely separated countries as California, Alaska, and Europe, rather than to any real difference in the date of the disturbances in the several regions."

This, of course, may be true, but the evidence of a revolution affecting the entire area from Mexico to Alaska at the same time is meager.

Crickmay in 1931 attempted to show that the "Jurasside" (Cordilleran, Nevadian, Sierra Nevada, Jurassic) orogeny of previous writers occurred at the close of the Jurassic and was of little importance when compared with an earlier diastrophism which he called the Agassiz orogeny. He made the Nevadian revolution a many phase disturbance compounded of the Agassiz, "Jurasside," and "mid-Cretaceous" compressions. Crickmay based the Agassiz orogeny, which he placed between the Callovian and Argovian (lower Oxfordian), on a thick basal conglomerate of the Agassiz Series (Argovian) in the Harrison Lake region in British Columbia and cited as corroborative evidence for California "certain coarse sediments of the Franciscan Series." He maintained that this "orogeny" was of great magnitude and wide extent, was accompanied by batholithic intrusions, and affected the Cordilleran region from California and Nevada to central Alaska. The writer does not doubt that there was a disturbance at the time of the Agassiz "orogeny" or that it might have been widespread but a major orogeny, of the magnitude and extent postulated, could not be created out of an unconformity and a basal conglomerate in one locality and the presence of sandstone at a few localities, especially when the coarse Franciscan sandstones referred to are later than the Argovian and actually (in Oregon) rest unconformably on Kimmeridgian sediments and volcanics. The presence of granitic pebbles in the basal Agassiz conglomerate is not proof that the "Agassiz orogeny" was accompanied by batholithic invasion. Granitic pebbles are not uncommon in the Amador and Mariposa but this cannot be used as evidence of early Upper Jurassic batholiths since they were clearly derived from

an early Paleozoic or pre-Cambrian crystalline complex to the west. As pointed out by Hinds (1935), particularly for northern California, "the field evidence upon which this Agassiz orogeny is based seems highly insecure." There is no evidence anywhere in California for orogeny and batholithic intrusion at the supposed time of the Agassiz orogeny. Warping, tilting, and advances and retreats of the sea, resulting in unconformities, disconformities, and overlaps undoubtedly occurred at various times during the Jurassic, but they are of very minor importance when compared with the movements which deformed the Mariposa.

Regarding the Jurasside (Nevadian) orogeny Crickmay (1931, pp. 59, 63) states:

> "A tentative interpretation may be made of the universal unconformity at the base of the Cretaceous rocks. The great extent of this structure makes it seem likely that mountain structures were formed over all the country between the ocean and the site of the present Sierra Nevada, Klamath, and Cascade Mountains. The folding may have been initiated by the Agassiz orogeny or even earlier, and augmented by the Jurasside. Some was certainly compressed still closer by the mid-Cretaceous (pre-Chico) compression. . . . The Jurasside diastrophism was somewhat overestimated by the older geologists. It was hardly a revolution, but rather a disturbance. Its importance has been exaggerated, partly, no doubt, because the effects of the Agassiz Orogeny have been confused with it."

The above statement ignores practically all that is definitely known regarding the geology of the Sierra Nevada and the Coast Ranges of California. It confuses disconformities and slight unconformities caused by warping or mild synorogenic disturbances with an orogeny such as that which deformed the Mariposa and older beds and resulted in tight folding, overturning, thrust faulting and the formation of dynamically metamorphosed rocks over wide areas in the Sierra Nevada.

In 1935 Hinds reviewed the ideas that had been advanced regarding the Nevadian revolution, suggesting the term Nevadan used by the writer. He concluded that the folding, faulting, and dynamic metamorphism and the batholithic invasions occurred in the late Upper Jurassic, before the deposition of the Upper Jurassic Knoxville.

### WESTERN BELT OF JURASSIC ROCKS

In discussing the Nevadan orogeny the writer will take up the observed effects in California and southwestern Oregon; this does not imply, however, that a much larger region was not affected.

In the Sierra Nevada and northwestward into southwestern Oregon the Mariposa and older beds have been strongly deformed, converted into slates and locally into schists, and extensively invaded by serpentines and granodiorite. The belt of deformed Jurassic rocks cannot be traced continuously from the Sierra Nevada into southwestern Oregon because of subsequent erosion and because of a broad belt of Tertiary

volcanics and sediments which cover the older structures at the northern end of the Sierras and the Sacramento Valley, a belt more than 70 miles wide. Immediately northwest of this belt the younger Jurassic sediments (Mariposa, Monte de Oro, Amador) have been removed by erosion but on the Trinity River the Mariposa again appears and continues to the north into southwestern Oregon. Because of these interruptions it cannot be stated positively that Upper Jurassic sediments were deposited in a continuous belt from the Sierra Nevada, through northern California into southwestern Oregon. However, the nature of the sediments and the faunas leads to the belief that a continuous Upper Jurassic sea existed in this region, a sea which transgressed from the east onto a land area occupied by the present central Coast Ranges and present margin of the Pacific. The constant westward coarsening of the Mariposa and Amador, as well as the character of the debris clearly shows a source to the west.

#### EASTERN SIERRAN BELT OF TRIASSIC AND JURASSIC ROCKS

The Mother Lode thrust is the eastern boundary of the Mariposa and associated beds; east of this thrust they have been removed by erosion and a broad belt is occupied only by Paleozoic sediments and volcanics and the plutonics which accompanied the Nevadan revolution. Twenty-five to 50 miles east of the western belt of Upper Jurassic are discontinuous areas of Upper Triassic and Jurassic sediments and volcanics; doubtless these formed a continuous belt at one time but since they lie in the region of maximum plutonic invasion and maximum Tertiary uplift they have been obliterated in many places. South of Taylorsville, near the northern end of the Sierras, these beds have been named the Milton and Sailor Canyon formations but there is little doubt that they are equivalent; since the name Milton has priority it will be used here to include all the Triassic and Jurassic rocks of the higher parts of the Sierra Nevada, south of Taylorsville. Undoubtedly these beds could be divided into several distinct formations. Practically all known occurrences have been visited by the writer, but much of the following information regarding the details of these beds has been obtained from Clark (1930).

Where the Milton sediments and volcanics have not been engulfed by plutonic intrusions or removed by erosion they lie in a broad steep-limbed syncline, practically free from minor crumpling, thrusting, and overturning. As in the case of the Upper Jurassic sediments to the west the evidence is clear that the Milton was derived from the west. The conglomerates contain abundant debris of the Paleozoic rocks (Calaveras) and thicken and coarsen westward.

The best exposed and most complete section is on the north fork of the American River, in Placer County. On the west limb of the syncline, which dips 50° to 60° E., basic and intermediate volcanics and radiolarian cherts, 200 feet thick, are followed by 12,800 feet of conglomerates, sandstones, hard slaty shales, and fine-grained andesitic tuffs. The center of the syncline is occupied by 9500 feet of intermediate and basic volcanics, flows, agglomerates, and tuffs, with thin interbedded sediments. Only about 9000 feet of sediments and tuffs lies below the volcanics on the east limb of the syncline, the lower part having been obliterated by batholithic intrusions. Well-preserved Upper Triassic fossils are found at and near the base of the sediments on the west limb of the syncline, Lower Jurassic fossils 2500 feet above the base, and Middle Jurassic fossils 9500 feet above the base; no fossils have been found in the upper 13,000 feet of sediments and volcanics. The section is well exposed, and no unconformities or disconformities have been observed. No sediments of known Mariposa age have been found in this section, but possibly a part of the unfossiliferous upper 13,000 feet may be equivalent to the Mariposa. The upper volcanics are possibly equivalent to the extensive Logtown Ridge volcanics lying between the Amador and Mariposa west of the Mother Lode thrust. No evidence supports the contention that the Mariposa and Milton were deposited in separate basins or that they were separated by a period of orogeny and batholithic invasion. Observations of the thick and rather complete Jurassic section of the Taylorsville region give no indication of such a separation.

In the Taylorsville region in Plumas County near the northern end of the Sierra Nevada are fossiliferous Lower, Middle, and Upper Jurassic sediments and volcanics. The uppermost formation, the Foreman, is generally considered equivalent to a part if not all of the Mariposa. The highest known Jurassic in the Taylorsville region is either upper Oxfordian or lower Kimmeridgian. Much of the Taylorsville Jurassic is similar lithologically to the Milton and lies along the same general trend, 23 miles north of the most northerly exposure of the Milton. The Taylorsville Jurassic and the Milton are obliterated by batholithic intrusions and covered by Tertiary volcanics in the 23 miles separating them, so that they cannot be traced across this gap.

In the Taylorsville region the Paleozoics are thrust eastward over the Jurassic overturning it toward the east—just the opposite of conditions along the Mother Lode thrust on the west flank of the Sierra Nevada. The Taylorsville region is east of the main axis of the Sierra Nevada and the Jurassic section is lower both topographically and structurally than the Milton, a fact which may account for the preservation of the Upper Jurassic in this region.

### COMPARISON OF EASTERN AND WESTERN BELTS

The Mariposa, Amador, and Monte de Oro beds, on the west flank of the present Sierra Nevada, are much more intricately folded, thrust-faulted and dynamically metamorphosed than the Milton and the Taylorsville Jurassic. The structure of the Milton, a broad steep-limbed but rather symmetrical syncline, is much simpler than that of the Mariposa, which is nearly always overturned and closely folded. The Milton, however, has been more extensively intruded and more frequently obliterated by batholithic intrusions. The Mariposa has been intruded by batholithic bodies and occasionally obliterated but not on the same scale as the Milton. In general the Milton shows a greater degree of thermal metamorphism than the Mariposa but less dynamic metamorphism. This is the reason for the better preservation and general lack of distortion of the fossils in the Milton.

Possibly several advances and retreats of the sea and local warping, uplift, and depression occurred during the deposition of the Taylorsville Jurassic, the Milton, and the Mariposa, such as those described by Lupher (1941) for the Jurassic of east central Oregon, but the evidence is definitely against intense folding or batholithic intrusion in the Sierra Nevada earlier in the Jurassic than the strong orogeny which folded and compressed the Mariposa. There is no evidence for the "Agassiz orogeny" or that the Nevadan orogeny was a many phase event.

In the Sierra Nevada the maximum effect of the Nevadan folding, resulting in overturned folds, often of great amplitude, great thrusts, such as the Mother Lode zone, and the widespread development of dynamically metamorphosed rocks, lies along a belt on the west flank of the present mountains both to the east and west of the Mother Lode thrust. The eastern belt of Triassic and Jurassic rocks, near the present crest of the Sierras, are strongly folded, but the great syncline in which they lie is much less compressed and the rocks are less dynamically metamorphosed than those in the belt along the present western flank of the mountains. The Taylorsville Jurassic is strongly folded and thrust-faulted perhaps representing an eastern belt of strong deformation formed during the Nevadan orogeny. However, since this east-vergent zone is only exposed in one comparatively small area no broad generalization can be drawn. Its possible continuation to the southeast, along the eastern margin of the ancestral late Jurassic Sierra Nevada, is lost beneath Tertiary lavas and obscured by late Tertiary and Pleistocene block faulting. More will have to be known regarding the structure of the region east of the present Sierra Nevada before any conclusions regarding this eastern zone of folding can be reached. The only state-

ment that can be made is that the intensity of the deformation decreases eastward in the present Sierra Nevada.

The western belt of maximum Nevadan deformation continues northwesterly along the western flank of the Sierras until it disappears beneath Tertiary sediments and volcanics. It reappears again in the northern Coast Ranges of California and continues to the north and north-northeast into southwestern Oregon. In northern California and southwestern Oregon the conditions are very similar to those along the western flank of the Sierras. Paleozoic beds are thrust westward over the Jurassic and the axial planes of the folds in the Jurassic dip eastward. The belt is quite as wide in Oregon as in the Sierra Nevada, but the folding is less intense and possibly may be dying out. The northward continuation is buried beneath Tertiary sediments and volcanics near Roseburg, Oregon.

### DATE OF NEVADAN OROGENY

The youngest beds deformed by the Nevadan orogeny, both in California and southwestern Oregon are Kimmeridgian, thus establishing a definite lower limit for the orogeny.

In the Sierra Nevada the oldest undeformed beds resting on the Mariposa are Upper Cretaceous, and there is little direct evidence regarding the upper time limit of the Nevadan orogeny. However, in northern California unmetamorphosed Franciscan is in close proximity to metamorphosed Mariposa and in Oregon unmetamorphosed Franciscan and Knoxville rest unconformably on Galice slates. Furthermore, the gently dipping Upper Cretaceous sands and shales, resting on closely folded dynamically metamorphosed Mariposa on the east side of the Sacramento Valley, reappear on the west side of the Valley resting disconformably on Lower Cretaceous beds. These, in turn, rest with little or no angular discordance on unmetamorphosed Upper Jurassic Knoxville sediments which grade downward into the Franciscan without break. In the Coast Ranges of California and Oregon the Mariposa was strongly deformed and dynamically metamorphosed before the deposition of the Franciscan and Knoxville. The upper part of the Knoxville is Tithonian (Aquilonian) and the lower part of the Knoxville and the upper part of the Franciscan are upper Portlandian. The age of the lower part of the Franciscan is uncertain, although it is younger than the lower Kimmeridgian. The coarse arkosic sediments which predominate in the lower part of the Franciscan probably accumulated rapidly and could have formed within the limits of Portlandian, or even a part of Portlandian time. Thus, a fairly definite upper time limit can be established for the Nevadan orogeny.

The writer believes there is definite and reliable evidence that the Nevadan orogeny took place immediately or very shortly after the deposition of the Mariposa (upper Oxfordian and lower Kimmeridgian) and before deposition of the Franciscan. The ancestral Sierras were uplifted at that time and the upper, little metamorphosed sediments quickly removed; many events were crowded in what appears to be a comparatively short time. In commenting on Smith's (1909; 1910) dating of the Nevadan revolution Knopf (1929, p. 14) states:

> "If these conclusions are true, then one of the mightiest of physical revolutions, one that caused the close folding of the strata of the Sierra Nevada and the subsequent intrusions of immense batholithic masses, beginning with peridotite and ending with enormous volumes of granite, ran its course during a fractional part of late Jurassic time."

The writer sees no escape from the conclusion that the Nevadan folding, with uplift and consequent erosion, took place with great rapidity. As has been pointed out by the writer (1941) the apparent rapidity of diastrophism has been characteristic of the Cretaceous and Tertiary history of the Coast Ranges. In general, even severe diastrophisms occupy a relatively short space of time when compared with the intervening periods of deposition. This seems particularly true of the mobile belts bordering the Pacific.

### EFFECT OF THE NEVADAN OROGENY WEST OF THE SIERRA NEVADA

The present Coast Ranges and the region bordering the present coast stood as a land mass of rigid crystalline rocks during the deposition of the Monte de Oro, Amador, and Mariposa; it was the foreland of the Upper Jurassic sea which lay to the east. During the Nevadan orogeny the Sierran rocks were folded, crushed, and thrust against and over this foreland. The effect on the foreland, the present Coast Ranges, is not known, but as these rocks already had experienced severe dynamic metamorphism, had been intruded by batholithic bodies, and eroded to a sufficient depth to expose the plutonics it is not likely they underwent any severe folding or further metamorphism.

The ultimate effect of the Nevadan orogeny on the region of the present Coast Ranges was of great and far reaching importance, an effect that, heretofore, has not been recognized, chiefly because of a lack of knowledge of the age of the Franciscan and because of the various erroneous correlations. A direct result of the severe folding and compression of Mariposa and older Sierran rocks was the uplift of the Cordilleran region as a whole. Eastern California, eastern Oregon and Nevada were never again invaded by the sea. The ancestral Sierra Nevada probably did not attain any great elevation at this time but probably rose as a low land mass which was quickly stripped of the

unmetamorphosed surface beds. The foreland to the west of the ancestral Sierra Nevada was depressed and flooded by the sea, probably for the first time in the Mesozoic. The region west of this depressed belt, a comparatively short but unknown distance west of the present coast line, was a high and rugged land mass which appears to have been a rejuvenated and uplifted remnant of the former land mass. In this long geosyncline between the newly risen Sierra Nevada and the rugged land mass to the west the Franciscan and Knoxville were deposited. The only known place where this geosyncline spread onto the folded Jurassic rocks of the ancestral Sierras is in northern California and southwestern Oregon.

The Nevadan orogeny was of great importance in the geologic history of the Coast Ranges since it was the indirect, if not the direct cause of the long enduring upper Mesozoic geosyncline in which the thick upper Upper Jurassic and Cretaceous sediments of the California Coast Ranges and of southwestern Oregon accumulated. Aside from the ancient crystalline schists and plutonics exposed here and there in the Coast Ranges, the details of whose history are little known, the first very important decipherable event in the history of the Coast Ranges is the development of this broad depression in the upper Upper Jurassic which continued to sink and receive sediments during the remainder of the Jurassic and the Cretaceous. Fully 25,000 feet of upper Upper Jurassic sediments and a maximum of 55,000 feet of Cretaceous deposits accumulated; the maximum known thickness of sediments in any one locality is slightly over 40,000 feet. The complex history of this geosyncline has been traced by the writer in a paper now in press (1941).

### THE BATHOLITHS

The time of emplacement of the granodioritic batholiths of the Sierra Nevada and adjacent regions has been stated as Upper Jurassic, Lower Cretaceous, and even Upper Cretaceous by various writers. The great granodioritic bodies are later than the folding and thrusting, known to have taken place either in the late Kimmeridgian or early Portlandian, thus fixing a definite lower time limit for the intrusions. The upper time limit is somewhat uncertain and the various statements that have been made are little better than guesses.

No evidence exists of granodioritic debris having an eastern source in the Franciscan, Knoxville, or Lower Cretaceous, but this does not mean that the batholiths were not emplaced at an early date; either they had not been stripped of their overlying cover or they were not in the drainage area of the late Upper Jurassic and Lower Cretaceous seas. The earliest sediments definitely known to contain granitic debris

from an eastern source are of Upper Cretaceous age, hence the only positive statement that can be made is that they are post-early Kimmeridgian and pre-Upper Cretaceous. However, Upper Cretaceous rocks, on the east side of the Great Valley of California, rest on a surface cut into highly folded bed-rock and granodiorite indicating that they must have been intruded and deeply eroded before the beginning of the Upper Cretaceous. The writer favors a late Upper Jurassic age for the emplacement of the batholiths but admits that evidence for this conclusion is not positive.

The batholiths do not represent a single, simple uprise of magma. Some bodies were emplaced before others, so that sometimes one granitic body intrudes another. In southern Mariposa County, California, the writer has mapped a complex batholithic body intrusive into the Mariposa and Amador that varies from a quartz-rich fine-grained granodiorite to a coarse-grained augite diorite. Intrusive into this is a coarse-grained biotite hornblende granodiorite similar to the great bulk of the Sierran granodiorites. The contacts between the two are clean cut, and there is some indication that the earlier body had completely solidified and perhaps cooled somewhat before the intrusion of the later granodiorite. The complete emplacement of all the batholithic masses may have occupied a fairly long time, possibly beginning in the late Upper Jurassic and continuing into the Lower Cretaceous.

The entire sequence of Nevadan orogeny and batholithic emplacement may be spoken of as the Nevadan revolution, but it must be remembered that the final emplacement of the batholiths may have occurred much later than the folding.

## THE DIABLAN OROGENY

The Franciscan and Knoxville were deposited in a geosyncline resulting from the Nevadan orogeny and continuing to receive sediments throughout the remainder of the Jurassic. Minor movements, resulting in local disconformities, occurred during the deposition of the Franciscan and Knoxville, especially in the western part of the present Coast Ranges. Since these become more pronounced westward, they probably represent strong movements in the land mass and in that part of the geosyncline west of the present coast line. However, none of these movements was severe or extensive in the present Coast Ranges. No break occurs between the Franciscan and Knoxville and deposition was essentially continuous throughout the Portlandian and Tithonian.

At the close of the Jurassic the entire Coast Ranges were uplifted and slightly folded. As in the case of the earlier minor movements and practically all succeeding Cretaceous and Tertiary movements, the

intensity of this diastrophism increased westward. Along the west side of the Sacramento Valley the base of the Lower Cretaceous is marked by a conglomerate, but only locally is there a noticeable angular discordance. However, northward the base of the Lower Cretaceous gradually transgressed at least 25,000 feet of Franciscan and Knoxville beds and, in Tehama County, California, rests on Paleozoic rocks. Westward, within the Coast Ranges, the Lower Cretaceous basal conglomerate (base of the Paskenta stage of the Shasta) contains abundant debris of both Franciscan and Knoxville and frequently overlaps the Knoxville and rests on the Franciscan.

The Diablo Range, which lies along the west side of San Joaquin Valley, is made up of a core of Franciscan rocks, flanked by Cretaceous sediments. Because of faulting, Knoxville beds are only occasionally exposed on the east side of the range, but on the west side Knoxville sediments are at least 3000 feet thick. Knoxville sediments are exposed in a number of places within the range, infolded into the Franciscan. Shasta (Lower Cretaceous) beds are occasionally exposed on both sides of the range and here and there within the range. On the two sides the Lower Cretaceous rests on the Knoxville either disconformably or with slight angular discordance. Within the range the Lower Cretaceous frequently transgresses the Knoxville on to the lower part of the Franciscan, indicating uplift and erosion between the Upper Jurassic and the Lower Cretaceous. In the Santa Lucia Range the structure is more complex, but Lower Cretaceous frequently laps over the Knoxville onto the Franciscan. In both the Diablo and Santa Lucia Ranges, as well as in the northern Coast Ranges, occasional breccias at the base of the Lower Cretaceous contain angular blocks of Franciscan and Knoxville up to 10 feet in diameter.

Evidence of uplift, folding, faulting, and erosion is abundant throughout the Coast Ranges between the Upper Jurassic and the Lower Cretaceous. The writer has named this diastrophism the Diablan orogeny from the Diablo Range. The effect of this orogeny is not comparable to the preceding Nevadan but it was widespread and resulted in strong unconformities and overlaps. It was stronger in the west and died out eastward toward the present eastern edge of the Coast Ranges. This orogeny did not destroy the Mesozoic geosyncline, as the Lower Cretaceous is deposited over approximately the same area as the Franciscan and Knoxville. It may have caused a slight eastward shifting of the main axis of the geosyncline but the evidence for this is not conclusive.

Like the more severe and widespread Nevadan orogeny the diastrophism at the close of the Jurassic must have occupied a comparatively

short interval as the upper Knoxville contains a very late Upper Jurassic fauna (late Tithonian) and the basal lower Cretaceous conglomerate represents a very low stage (Valanginian) in the Lower Cretaceous.

One reason for the widespread belief in an unconformity between the Franciscan and Knoxville is that the unconformity at the base of the Lower Cretaceous, the result of the Diablan orogeny, was erroneously believed to be at the base of the Knoxville, which was originally placed in the Lower Cretaceous.

## PROPOSED CORRELATIONS

The correlations between the Jurassic of southwestern Oregon and California proposed by the writer after many years of work in the Sierra Nevada and Coast Ranges of California and a limited amount of time in southwestern Oregon are embodied in Table 1:

TABLE 1.—*Proposed Jurassic correlations*

| | European Stages | California | Southwestern Oregon |
|---|---|---|---|
| Lower Cretaceous | Lower Albian<br>Aptian<br>Barremian<br>Hauterivian<br>Valanginian | Shasta group<br>Horsetown stage<br>Paskenta stage | Shasta group (restricted Myrtle)<br>Horsetown stage<br>Paskenta stage |
| | Unconformity | | |
| | Diablan orogeny | | |
| Upper Jurassic | Tithonian and Portlandian | Knoxville and Franciscan | Dillard (restricted)<br>Knoxville and Franciscan |
| | Profound Unconformity | | |
| | Nevadan Revolution | | |
| | Lower Kimmeridgian and Upper Oxfordian | Mariposa group | Galice group |
| | Older than Upper Oxfordian, may extend downward into Middle Jurassic | Amador group<br>Monte de Oro beds | Dothan group<br>Plant beds of Douglas County, Oregon |

The Amador group and the Monte de Oro beds and their apparent equivalent in Oregon, the Dothan group and the "Plant beds of Douglas County" have not been assigned to a more definite age than "older than upper Oxfordian" because so little is positively known regarding their age. Although the Amador and Dothan are older than the Mariposa and Galice they are closely related, and it is doubtful if they extend downward into the Middle Jurassic. The flora of the Monte de Oro beds

and the "Plant beds of Douglas County" has been correlated with the Middle Jurassic flora of England and Siberia, but the flora of both these units should be restudied before a positive statement as to their age is made.

## RÉSUMÉ OF JURASSIC HISTORY

So much of the evidence has been obscured by strong folding and thrusting, obliterated by batholithic invasion, and removed by erosion that only an incomplete record remains of the undoubtedly complex Jurassic history of California and southwestern Oregon. However, a brief summary, incomplete as it must be, and an interpretation of the history will be of value.

The Upper Triassic sea encroached from the east on a land mass which occupied a large part of central and northern California and southwestern Oregon. How far to the west, beyond the present coast, this land mass extended is, of course, unknown. It may have been essentially the same as the hypothetical Paleozoic land mass of Cascadia. However, it seems to have extended farther east than Cascadia, as Paleozoic sediments are found fully 60 miles west of the most westerly Triassic rocks in the Sierra Nevada. Figure 2 shows the approximate location of the western limit of the Upper Triassic sea. This is based on the present distribution and thickness of the Upper Triassic sediments; because of deep subsequent erosion it is approximate both in form and position and necessarily ignores any minor irregularities which must have existed. From the meager evidence available it is thought that the Triassic sea gradually encroached on California from the east; only the extreme eastern part was reached in the Lower Triassic. The entire Triassic section appears to thicken eastward. No break in sedimentation exists between the Upper Triassic and Lower Jurassic in western Nevada and eastern California and the reported unconformity in the Redding district in northern California is so slight that it is, at most, a disconformity. Sinking continued and the Lower Jurassic sea encroached westward on the western land mass, but not as far west as the present Mother Lode district of the Sierra Nevada. The position of this shore line is not known away from the Sierra Nevada as there are no known Lower Jurassic beds in either southern California or southwestern Oregon. This westward migration of the Jurassic sea continued into the Middle Jurassic; minor warpings and fluctuations occurred but there is no definite evidence of widespread movement. If the paleobotanical evidence of the age of the Monte de Oro and Bucks Peak beds is accepted, two widely separated points that must have been very close to the margin of the Middle Jurassic sea are established. Aside from these two points the exact position of its western

edge is very indefinite. Indications of the westward extent of the Upper Jurassic sea are fairly good, but only meager evidence exists for events which may have occurred toward the close of the Middle Jurassic and the beginning of the Upper Jurassic. The sea in which the Amador and Mariposa were deposited clearly transgressed from the east onto the old land mass which existed in the Upper Triassic and Lower and Middle Jurassic. However, the Nevadan revolution and subsequent erosion have obliterated much evidence so that it is not known whether the Amador, Mariposa, and Galice were deposited in the same general trough as the Upper Triassic and Lower and Middle Jurassic or in a separate basin to the east of that containing the earlier Mesozoic beds. The earlier Mesozoic sinking and westward encroachment may have continued with little or no interruption into the Kimmeridgian or a barrier may have been raised somewhere east of the present Mother Lode prior to the deposition of the Amador. The only evidence available is the presence of Upper Triassic, Lower, Middle and Upper Jurassic beds in the Taylorsville region in the northern Sierras. Although it is unlikely that all pre-Upper Kimmeridgian stages are represented in the Jurassic of Taylorsville there is no evidence of any important unconformity in the section and certainly no evidence of any extensive orogeny. Several minor advances and retreats of the sea may have occurred in the Taylorsville region but evidently the beds were not strongly folded prior to the Nevadan orogeny. Furthermore, the Milton sediments and volcanics are less severely folded and less metamorphosed dynamically than the Amador and Mariposa and this is believed to be reasonable evidence that they were not severely folded before the deposition of the Upper Jurassic. Also Amador and Mariposa sediments and volcanics, 10,000 to over 20,000 feet thick, end abruptly against the Mother Lode thrust and 20 miles to the east, near the crest of the Sierra Nevada, are Lower and Middle Jurassic beds over 20,000 feet thick. Even though this present 20-mile separation was greater before the Nevadan orogeny it still greatly limits the width of the marine basin in which the more than 20,000 feet of Amador and Mariposa was deposited. Although the evidence is far from positive it is believed that the Upper Jurassic sediments and volcanics west of the Mother Lode accumulated in essentially the same basin as the older Jurassic beds to the east and in the Taylorsville region. The sea transgressed westward and the basin expanded in that direction. The main axis of the subsiding trough and the zone of maximum accumulation of sediments may have shifted westward although no evidence was observed that such changes in configuration resulted from severe folding. In the region to the east of California diastrophisms at various times during the Jurassic probably caused warping and basin shifting in what is now

the Sierra Nevada. Many minor changes probably took place and the main axis of the geosyncline gradually shifted westward but no profound changes or basin fragmentation took place prior to the Nevadan orogeny in the Sierran region.

The final collapse of the Sierran geosyncline occurred after the deposition of the Mariposa, probably in the late Kimmeridgian, and the entire Sierran region was intensely folded and, along its western margin, thrust westward against and over the old foreland (or the rigid foreland was thrust downward beneath the basin) and the ancestral Sierra Nevada started to rise. The thrusts and axial planes of the overturned folds now dip 45° to 75° E., but these are only the roots of the old thrusts and folds whose upper parts have been removed by erosion. They probably decreased in angle upward and formed relatively low-angled thrust sheets.

As the ancestral Sierras rose the old foreland was depressed and invaded by the sea for the first time in the Jurassic. At the same time a part of the western land mass was uplifted and contributed great quantities of coarse debris to the geosyncline which continued to sink and receive sediments throughout most of the remainder of the Mesozoic. It was in this geosyncline, a result of the Nevadan orogeny, that the Franciscan and Knoxville were deposited in the Portlandian and Tithonian. The eastern limit of this geosyncline is fairly well known but its western limit, west of the present coast line, is conjectural. The approximate position of the Franciscan-Knoxville geosyncline is shown in Figure 3.

Minor movements during the deposition of the Franciscan and Knoxville resulted in local disconformities. These may have been severe in the region west of the present coast, but their effect was slight in that part of the geosyncline now occupied by the Coast Ranges.

At the close of the Jurassic rather widespread uplift and erosion occurred before the Lower Cretaceous sea spread over the region. Along the west side of the Sacramento and San Joaquin valleys the late Upper Jurassic beds were little disturbed, but westward, especially in the Diablo and Santa Lucia Ranges and in parts of the northern Coast Ranges, folding was strong and much of the Knoxville was removed before the deposition of the Lower Cretaceous. Although the Diablan orogeny, at the close of the Jurassic, was widespread it did not destroy the geosyncline in which the Franciscan and Knoxville were deposited and the Lower Cretaceous sea again flooded the same general area.

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# STRATIGRAPHY, STRUCTURE, AND PETROLOGY OF THE MT. CUBE AREA, NEW HAMPSHIRE

BY JARVIS B. HADLEY

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## ABSTRACT

Seventy per cent of the bedrock in the Mt. Cube area consists of metamorphic rocks of sedimentary and volcanic origin, ranging from middle Ordovician (?) to early Devonian. The total thickness is about 16,000 feet. Two of the oldest stratigraphic units, the Orfordville formation and the Piermont member of the Albee formation, are described in detail because this is the type locality.

The Orfordville formation, presumably middle Ordovician, is 5000 feet thick, and consists of interbedded black schist and volcanic rocks. The schist represents carbonaceous marine mud and sand. The volcanics represent metamorphosed basalt flows and pyroclastics (now amphibolite) with some rhyolite, quartz latite, and dacite (now biotite gneiss). The Piermont member, 1000 feet thick at maximum and locally absent, comprises most of the lower part of the Albee formation, and consists of aluminous, carbonaceous schist interbedded with noncarbonaceous quartz-mica schist, quartzite, and fine conglomerate.

About 30 per cent of the bedrock consists of gneissic igneous rocks, ranging from granite to quartz diorite. Most of them belong to the late Devonian Oliverian and New Hampshire magma series.

The stratified rocks were severely folded, broken by thrust faults, and metamorphosed during the late Devonian. Thrusting and overturning of folds is toward the east-southeast, except in the southeastern part of the area, where it is toward the west. The concordant, domical intrusives of the Oliverian magma series were active during the later stages of the folding, and represent the tops of syntectonic phacoliths, rather than pretectonic laccoliths.

The metamorphic and tectonic relationships of minerals in the middle-grade zone of metamorphism are briefly discussed, and an area of wholesale retrograde metamorphism associated with a major thrust fault is described.

## INTRODUCTION

About two thirds of the area shown on the Mt. Cube quadrangle of the U. S. Topographic Atlas is in west-central New Hampshire, and the rest lies in eastern Vermont (Fig. 1; Pl. 4). An area of about 150 square miles, mostly in New Hampshire, is the subject of the present study.

Little work on the bedrock geology of the New Hampshire portion of the Mt. Cube quadrangle has been done since the early surveys of Jackson (1844) and C. H. Hitchcock (1877). Recent studies in adjacent

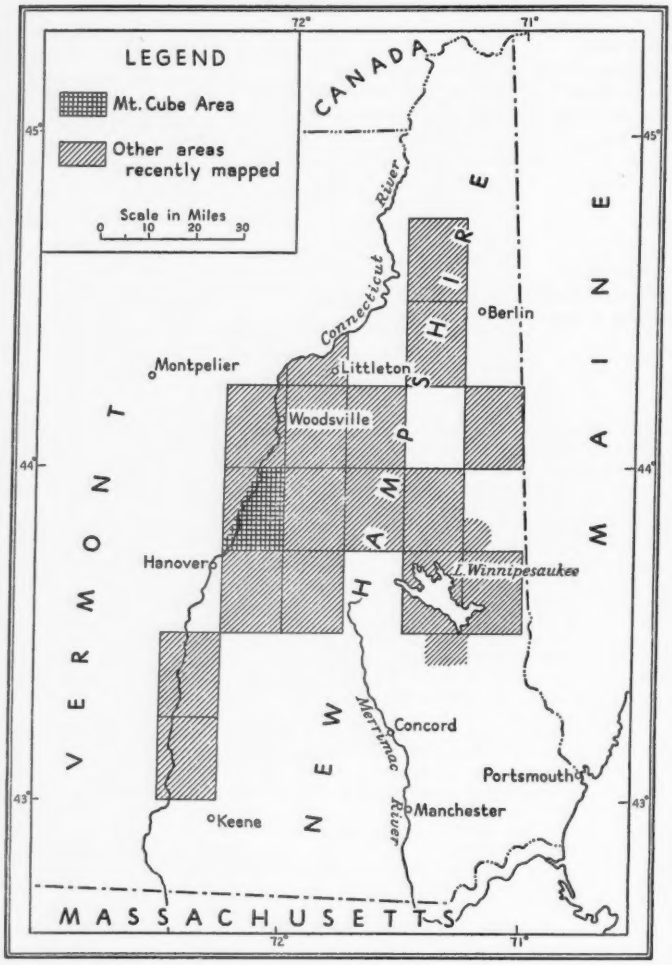


FIGURE 1.—*Index map*

regions, however, have contributed greatly to present knowledge of the geology of western New Hampshire. The work of Billings (1935; 1937) in the Littleton-Moosilauke area, adjoining the Mt. Cube area on the northeast, has been valuable in providing the basis for the stratigraphy and chronology of the present study.

The geological mapping in the Mt. Cube area occupied 29 weeks in the summers of 1934 and 1935. The excellent topographic map of the U. S. Geological Survey, published in 1928 on a scale of approximately a mile to the inch, was used as a base, outcrops being plotted on a three-fold, photostatic enlargement. In the higher and more heavily wooded parts of the area, an aneroid barometer, reading to 20 feet, was used. In addition, 54 miles of critical contacts in heavily wooded or structurally complicated areas were mapped by pace-and-compass traverse on a scale of 750 feet to an inch.

## ACKNOWLEDGMENTS

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Financial support for both field seasons was made available through a generous grant from the Milton Fund of Harvard University, and funds for thin sections and chemical analyses were supplied by the Department of Mineralogy and Petrography. The writer also expresses his gratitude to Professor J. W. Goldthwait and Mr. Frederic E. Everett who obtained funds for the publication of the colored geological map (Pl. 4).

## STRATIFIED ROCKS

### GENERAL STATEMENT

Seventy per cent of the bedrock in the Mt. Cube area consists of metamorphic rocks of sedimentary and volcanic origin, ranging in age from middle Ordovician (?) to early Devonian (Fig. 2). Many lithologic types are represented, and the total thickness is approximately 16,000 feet. The stratified rocks have the characteristics of middle-grade metamorphism, except in the northern part of the quadrangle, where rocks of low-grade occur.

Two of the lithologic units, the Orfordville formation and the Piermont member of the Albee formation, are more fully described than the other formations, because the Mt. Cube area is the type locality.

TABLE 1.—*Modes of the Orfordville formation*

| | | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Quartz | | 53 50–56 | 28.0 | 35 | 67 61–75 | 86 | 83 | 39.6 | 42 | 43 23–57 | 38 | tr 0– 1 | |
| Plagioclase | | | 25.4 | | | | | 14.8 | 42 | 35 20–45 | 53 | 28 10–44 | 20 |
| Orthoclase | | | | | | | | | 10 | 10 0–18 | | | |
| Muscovite | | 25 19–40 | 19.2 | 54 | 13 5–25 | 1 | 5 | 17.5 | 3 | 11 1–20 | 6 | | |
| Biotite | | 18 tr–22 | 23.4 | | 17 5–35 | 11 | | 19.3 | 3 | 1 0–2 | 3 | tr 0– 1 | 2 |
| Chlorite | | tr tr– 1 | | tr | tr 0– 2 | tr | | | | | tr | tr 0–tr | |
| Garnet | | tr 0–tr | | | 1 0–3 | 1 | | 2.2 | tr | tr 0–tr | tr | | |
| Staurolite | | | | | tr 0–tr | | | 4.1 | | | | | |
| Ottrelite | | | | 5 | | | | | | | | | |
| Amphibole | | | | | | | | | | | | 72 55–89 | 25 |
| Epidote | | | | | | | | | tr | tr 0–tr | tr | tr 0– 1 | 3 |
| Apatite | | | tr | | tr 0–tr | tr | | tr | | | | | |
| Sphene and leucoxene | | tr 0–tr | | | tr 0–tr | | | | | tr 0–tr | | tr tr | tr |
| Zircon | | | | | tr 0–tr | | | | | | | | |
| Magnetite and ilmenite | | 1 tr–2 | 1.8 | 2 | tr 0–tr | tr | | 0.5 | tr | tr 0–tr | tr | tr 0–1 | tr |
| Pyrite and pyrrhotite | | tr 0–tr | 0.8 | | tr 0–tr | tr | | 0.3 | | | tr | | tr |
| Calcite | | | | | | | | | | tr 0–tr | | tr 0–tr | 50 |
| Graphite | | 3 1– 6 | 0.9 | 4 | 2 tr– 2 | 1 | 12 | 1.0 | tr | tr 0–15 | | | |
| Totals | | 100 | 99.5 | 100 | 100 | 100 | 100 | 99.3 | 100 | 100 | 100 | 100 | 100 |
| Character of plagioclase | | oligoclase -andesine | $Ab_{69}$ $An_{40}$ | oligoclase -andesine | oligoclase -andesine | oligoclase -andesine | | $Ab_{65}$ $An_{35}$ | albite | oligoclase | oligoclase -andesine | andesine -labradorite | labradorite |
| Grain size mm. | Groundmass | 0.05 | | 0.03 | 0.1 | 0.1 | 0.1 | | 0.03 | 0.05 0.01–0.1 | 0.05 | 0.05–0.8 | 0.1–1.5 |
| | Porphyroblasts* | 1.00 | | 1.50 | 1.0 | | | | 1.0 | 1.0 0.1 –1.0 | 1.0 | | |

* Includes crystals in crystal tuff.

1. Mica schist, average and range of 5 sections.
2. Mica schist, calculated from chemical analysis in Table 2, column 1.
3. Mica schist with ottrelite, 1 section.
4. Quartz-mica schist, average and range of 9 sections.
5. Quartzite, average of 2 sections.
6. Black graphitic quartzite, 1 section.
7. Staurolite-mica schist, calculated from chemical analysis (Table 2, column 2).
8. Biotite gneiss: (soda-rhyolite) average of 2 sections.
9. Biotite gneiss: (quartz latite) average and range of 6 sections.
10. Biotite gneiss: (dacite) average of 3 sections.
11. Amphibolite, average and range of 7 sections.
12. Calcareous amphibole gneiss, average of 2 sections.

### ORFORDVILLE FORMATION

*General statement.*—About one third of the Mt. Cube area is underlain by dark-gray schists and metamorphosed volcanics, which constitute the oldest-known formation in western New Hampshire. To these rocks the

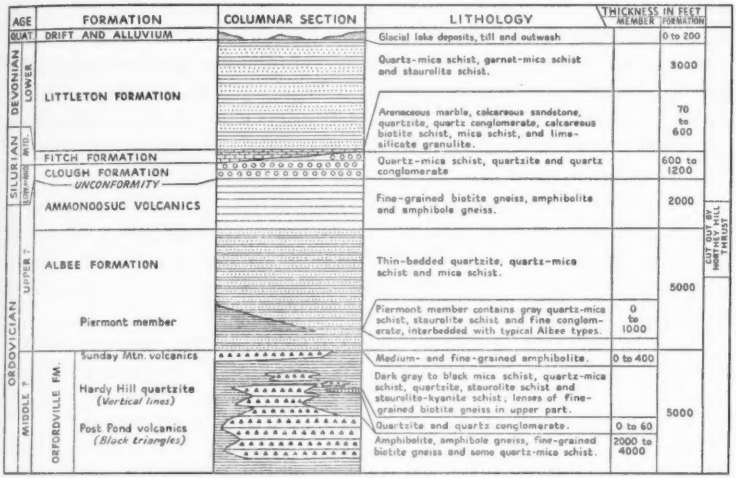


FIGURE 2.—*Columnar section for Mt. Cube area*

name Orfordville formation has been given (Hadley, 1938) after the village of Orfordville near the center of the Mt. Cube quadrangle. The formation represents about 5000 feet of more or less carbonaceous marine sediments, with interbedded flows and pyroclastics. The lower part consists of more than 2000 feet of submarine basalt flows, known as the Post Pond volcanic member. Midway in the sediments occurs a relatively thin, white quartzite and quartz conglomerate, the Hardy Hill quartzite (Chapman, 1939, p. 132). At the top of the formation is the Sunday Mountain volcanic member, essentially similar petrographically to the Post Pond member.

*Post Pond volcanic member.*—Besides a large body in the lower part of the Orfordville formation, two stratigraphically higher lenses south of Orfordville are included in the Post Pond member. In these lenses and throughout most of the main body, the dominant rock is dark-green, fine- to medium-grained amphibolite. It is more or less homogeneous, without banding or other conspicuous structure, except a moderately well-developed schistosity. However, at one place distorted pillows appear.

11. Amphibolite, average and range of 7 sections.
12. Calcareous amphibole gneiss, average of 2 sections.

5. Quartzite, average of 2 sections.
6. Black graphitic quartzite, 1 section.

Accompanying the amphibolite is also a large amount of banded hornblende gneiss, calcareous hornblende gneiss, and light-colored gneiss. The last is irregularly distributed throughout the main body of the Post Pond member, and occupies conspicuous areas to the east and southwest of Post Pond. Laboratory study shows that the light-colored gneiss includes: (1) quartz-albite-orthoclase gneiss (originally soda rhyolite), (2) quartz-oligoclase-orthoclase-muscovite gneiss (originally quartz latite), and (3) quartz-andesine-muscovite-biotite gneiss (originally dacite). Modes of rocks in the Post Pond member will be found in Table 1. At several places in the Post Pond member, there is carbonaceous schist similar to that characteristic of the main part of the Orfordville formation.

Amphibolite is generally composed of approximately 70 per cent hornblende and 30 per cent andesine ($Ab_{57}An_{43}$). Based on a chemical analysis of hornblende with similar optical properties from the Moosilauke quadrangle (Table 16), the chemical composition of the amphibolite is deduced as approximately that of basalt (Table 14).

Hornblende gneiss is similar to amphibolite, except that feldspathic layers, an inch to a foot thick, give the rock a banded appearance, and thin bands containing biotite are commonly present. The gneiss occurs in zones as much as 100 feet thick, associated with amphibolite.

Calcareous hornblende gneiss contains a large amount of calcite, both disseminated through the rock and in small pods and lenses of relatively pure marble. This is rather abundant locally in the amphibolite bodies, and forms pock-marked outcrops, due to weathering of the calcite. Some of the excessive calcite in these rocks may be due to precipitation of calcium carbonate from sea water as a result of eruptions on the sea floor. Some may represent amygdule fillings.

Quartz-oligoclase-orthoclase-muscovite gneiss (representing quartz latite) is the most abundant variety of the light-colored volcanics. Thin sections show an inequigranular mosaic of quartz and feldspar, with considerable muscovite and small amounts of biotite. Crystals of oligoclase 1 millimeter in diameter constitute as much as 20 per cent of some specimens; they are generally rounded and fragmented. Locally well-bedded varieties occur, with very thin gray layers containing opaque carbonaceous material. These features are thought to indicate a pyroclastic origin for much of the gneiss. Some of it is rather massive, however, and may represent flows.

Quartz-albite-orthoclase-gneiss (representing rhyolite) closely resembles the quartz-oligoclase-orthoclase-muscovite gneiss but is considerably less abundant. Locally it, too, has porphyritic phases and may be thin-bedded.

Quartz-andesine-muscovite gneiss (representing dacite or rhyodacite) commonly contains 5 to 15 per cent broken crystals of albite 1 millimeter across, in a moderately schistose matrix with a few per cent of biotite.

*Black schist.*—Most of the upper part of the Orfordville formation consists of dark-gray or black phyllitic schist. Quartz and mica predominate, and oligoclase-andesine is abundant in many specimens. The characteristic dark color is due to finely disseminated opaque material amounting to about 1 per cent. Chemical analyses (Table 2) show that this is carbon, and high magnification of a few specimens showed the presence of minute, hexagonal flakes of graphite. Syngenetic iron sulphide is disseminated widely in the darker schist, and its oxidation has given to weathered outcrops a characteristic brownish tinge.

Field and laboratory studies show that the ratio of quartz plus feldspar to mica varies considerably in the schists. Most of the schist, however, can be assigned to the following classifications: (1) quartz-mica schist, (2) mica schist, (3) quartzite, (4) carbonaceous quartzite. These terms have been arbitrarily used as follows: mica schist—less than 60 per cent quartz and feldspar, more than 40 per cent mica; quartz-mica schist—60 to 80 per cent quartz and feldspar; quartzite—more than 80 per cent quartz and feldspar. In addition, there are considerable areas in the western part of the formation where staurolite, or kyanite and staurolite are abundant in the schists, and near the eastern edge some ottrelite schist occurs. Modes of these types are given in Table 1.

Throughout the upper part, below the Sunday Mountain volcanic member, numerous small lenses of fine-grained, light-colored gneiss are interbedded with the schist. They represent original soda rhyolite, quartz latite and dacite pyroclastics, similar to those in the Post Pond member.

Quartz-mica schist and mica schist are most abundant. Bedding, represented by regular layers a fraction of an inch to several inches thick, is common but is generally obscured by the prominent schistosity. Porphyroblasts of biotite are characteristic; they are shaped like orange seeds or oat grains, 1 to 2.5 millimeters long, with their longest dimensions generally parallel to the schistosity. Garnet porphyroblasts from microscopic size to 2 millimeters are also abundant. Throughout the formation the schists are relatively fine-grained toward the east, but the recrystallization texture becomes coarser, with more abundant and larger porphyroblasts toward the west. Relatively coarse-textured schist is especially notable in a belt extending southward from Orford to the Ompompanoosuc River, between the Post Pond member and the Ammonoosuc thrust.

On the west slopes of Kenyon Hill and Strawberry Hill in Orford, unoriented porphyroblasts of staurolite, locally 2 to 3 inches long, occur in the mica schist. In the belt between the Post Pond member and the Ammonoosuc thrust, mica schist and quartz-mica schist contain por-

TABLE 2.—*Analyses of schists of the Orfordville formation*

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 61.52 | 66.26 |
| $TiO_2$ | .64 | .48 |
| $Al_2O_3$ | 17.59 | 16.54 |
| $Fe_2O_3$ | 1.54 | 1.42 |
| FeO | 4.86 | 3.53 |
| MnO | .04 | .09 |
| MgO | 2.76 | 2.76 |
| CaO | 2.04 | 1.04 |
| $Na_2O$ | 1.74 | 1.13 |
| $K_2O$ | 3.55 | 3.22 |
| $H_2O+$ | 1.96 | 1.92 |
| $H_2O-$ | 0.65 | 0.50 |
| $P_2O_5$ | tr | tr |
| $CO_2$ | none | none |
| C | .87 | .98 |
| S | .49 | .11 |
| | 100.25 | 99.98 |

1. Mica schist of the Orfordville formation, northeast ninth of Mt. Cube quadrangle. From 20 feet southeast of road, 750 feet N. 30° E. of road corner 1111, Clay Hollow. For mode *see* Table 1, column 2. W. H. Herdsman analyst.

2. Staurolite-mica schist of the Orfordville formation, central ninth of Mt. Cube quadrangle. From south summit of Blackberry Hill, 1¼ miles east of Orford. For mode *see* Table 1, column 7. W. H. Herdsman analyst.

phyroblasts of kyanite and staurolite an inch or more long, largely altered to muscovite and chlorite.

Beds of gray, micaceous quartzite, a foot or more thick, occur near the top and in the extreme western part of the formation; elsewhere the quartzite is not prominent.

Beds of coal-black, schistose quartzite, an inch to a foot or two thick, are associated with volcanic rocks throughout the formation. This rock is unusual in that the mica is entirely muscovite, and carbonaceous material amounts to 10 per cent in some thin sections. Locally the carbonaceous quartzite grades into well-bedded, gray gneiss of pyroclastic origin, or into black, muscovitic schist.

*Hardy Hill quartzite.*—This rock was named by Chapman (1939, pp. 132-133) in the Mascoma quadrangle, where it consists of "gray to white quartzite and quartz conglomerate, with minor amounts of interbedded quartz-mica schist and mica schist" and has a maximum thickness of 250 feet. In the Mt. Cube quadrangle, the Hardy Hill quartzite is

largely confined to a few thin remnants which lie 50 to 150 feet west of the Northey Hill thrust. The largest exposure, 100 feet wide, is in the extreme northeast part of the area. Along this belt, the member is believed to have been in part squeezed out and in part cut out by the thrust. The main belt of Hardy Hill quartzite in the Mascoma quadrangle, if traced northward toward the Mt. Cube quadrangle, would enter it approximately half a mile west of the Northey Hill thrust. The northward continuation of this belt in the Mt. Cube area has not been found, either because the quartzite is very thin and is not exposed or because it was not deposited in that part of the area.

*Sunday Mountain volcanic member.*—This name was given (Hadley, 1938) to amphibolites and interbedded fine-grained gneiss and black quartzite, constituting the top of the Orfordville formation. The member has a maximum thickness of 400 feet on Cottonstone Mountain and Sunday Mountain, and is not more than 200 feet thick northeast of Peaked Mountain. Between Sunday Mountain and Clay Hollow, it is too thin to appear on the geological map, or is absent. This is believed to be largely due to faulting or squeezing, although irregular deposition of the volcanics may have occurred. Petrographically the rocks of the Sunday Mountain member are similar to those previously described in the Post Pond member.

*Low-grade rocks of the Orfordville formation.*—In the western part of Piermont township, west of highway No. 10, the Orfordville formation is represented by rocks showing a lower grade of metamorphism. Here black slate is the metamorphic equivalent of the schist of the middle-grade zone; slaty felsites represent the interbedded volcanics; chlorite schist and greenstones, with black, schistose quartzite, represent the Sunday Mountain volcanic member; and sericitic quartz conglomerate 60 feet thick, with pebbles one half to three quarters of an inch long, is the Hardy Hill quartzite member. To the southwest in this same belt, low-grade equivalents of the Post Pond member were mapped during the summer of 1938 but are not shown on Plate 4.

*Thickness.*—Accurate measurements of thickness in such complexly folded rocks are, of course, impossible. An estimate of 5000 feet, with a probable error of 30 per cent, is obtained by comparing the areal extent of the Orfordville formation with that of the Albee formation, of which the thickness can be somewhat more accurately estimated. The Post Pond member is believed to represent volcanic lenses which may have an aggregate thickness of several thousand feet in the vicinity of Lyme and the south part of Orford. Yet it pinches out laterally at various

stratigraphic levels (Fig. 2), notably on Acorn Hill and north-northeast of Brick Hill in Lyme, and southwest of East Thetford, Vermont (Pl. 4).

*Age.*—As no fossils have been found in the Orfordville formation, its age cannot be definitely determined. It underlies the Albee and Ammonoosuc formations, which have a combined thickness of 7000 feet. These in turn are separated by an unconformity from the Clough formation and the fossiliferous middle Silurian Fitch formation. Thus the Orfordville formation is certainly pre-Silurian. In eastern Vermont and southern Quebec, the supposed Cambrian rocks are largely greenish-gray slates and quartzites. On the other hand, the Magog slates and associated volcanics (Clark, 1934, p. 11) of known middle Ordovician age are lithologically similar to the Orfordville formation. Therefore a middle Ordovician age is tentatively preferred.

### ALBEE FORMATION

*General statement.*—The Albee formation was originally described by Billings (1935; 1937) in the Littleton-Moosilauke area, where it is the lowest stratigraphic unit and its base is not exposed. In the Mt. Cube area about 2000 feet of the lower beds of the Albee formation is found immediately overlying the Orfordville formation in a belt northwest of the Northey Hill thrust. This belt has been traced by Billings from the north boundary of the Mt. Cube quadrangle, through the Woodsville quadrangle, into the Moosilauke quadrangle (unpublished map of the New Hampshire portion of the Woodsville quadrangle). Within the Albee formation, about 1000 feet above its base, is a lithologically distinct group of rocks named the Piermont member (Hadley, 1938).

*Lithology and petrography.*—The rocks in the Albee formation proper are dominantly light-gray, quartzitic, and moderately schistose. Stratification is more prominent than in the Orfordville formation; beds ranging from an inch or less to several feet thick are commonly seen, generally much contorted (Pl. 1, fig. 2). As in the Orfordville formation, fine-grained rocks predominate in the eastern part of the Albee formation, whereas recrystallization due to metamorphism has produced coarser textures in the western part. On a basis of field and laboratory study, the rocks of the normal Albee formation may be classed as follows: (1) quartzite, (2) quartz-mica schist, (3) mica schist. Modes estimated from thin sections of these rocks are given in Table 3. In most of the rocks, the microscope revealed oligoclase-andesine, similar to that in the Orfordville formation, amounting to as much as 30 per cent of some specimens.

FIGURE 1. THIN-BEDDED QUARTZITE OF CLOUGH FORMATION
Near the eastern summit of Mt. Cube. Beds strike nearly east-west, dip steeply to the southeast, and are isoclinally folded.

FIGURE 2. FOLDED QUARTZ-MICA SCHIST AND QUARTZITE IN ALBEE FORMATION
North slope, Peaked Mountain.

FIGURE 3. BIOTITE-QUARTZ DIORITE (LIGHTER COLOR)
Intruding hornblende-quartz diorite of Mascoma group on Holts Ledge cliffs.

OUTCROPS

FIGURE 1. SMARTS MOUNTAIN GRANODIORITE

FIGURE 2. GRANODIORITE OF MASCOMA GROUP
Biotite aggregates in well-developed foliation.

FIGURE 3. BETHLEHEM GNEISS FROM DEMMICK HILL (INDIAN POND PLUTON)
Linear aggregates of biotite on foliation surface.

FIGURE 4. QUARTZ MONZONITE GNEISS OF MASCOMA GROUP
One mile southwest of Sugar Hill, Mt. Cube quadrangle. Note lineation of biotite.

Quartzite is abundant, in massive beds or thinly interbedded with quartz-mica schist. It generally contains considerable lepidoblastic muscovite, which gives it a more or less weak schistosity. Numerous biotite porphyroblasts, a millimeter or less in diameter, are commonly

TABLE 3.—*Modes of the Albee formation*

| | | 1 | 2 | | 3 | 4 | 5 |
|---|---|---|---|---|---|---|---|
| Quartz, Plagioclase | | 88 | 69 | 60–75 | 49 | 38 | 65 |
| Muscovite | | 5 | 14 | 3–23 | 29 | 12 | 12 |
| Biotite | | 6 | 17 | 15–20 | 20 | 32 | 20 |
| Chlorite | | tr | tr | 0–tr | tr | | |
| Garnet | | 1 | tr | 0–1 | 1 | 7 | 2 |
| Staurolite | | | | | tr | 10 | |
| Apatite | | tr | tr | 0–tr | | | |
| Zircon | | tr | tr | 0–tr | | | |
| Tourmaline | | tr | tr | 0–tr | | | |
| Magnetite and ilmenite | | tr | tr | tr | 1 | tr | tr |
| Graphite | | | | | | 1 | 1 |
| Totals | | 100 | 100 | | 100 | 100 | 100 |
| Character of plagioclase | | oligoclase -andesine | oligoclase -andesine | | oligoclase -andesine | oligoclase -andesine | oligoclase -andesine |
| Grain size, mm. | Groundmass | 0.1 | 0.1 | .03–.15 | 0.03 | 0.05 | 0.05 |
| | Porphyroblasts | 1.5 | 1.0 | | 1.5 | 3.0–6.0 | 1.5 |

1. Quartzite, average of 2 sections.
2. Quartz-mica schist, average and range of 3 sections.
3. Mica schist, average of 2 sections.
4. Piermont member: staurolite-mica schist, average of 2 sections.
5. Piermont member: gray quartz-mica schist, 1 section.

scattered through the quartzite, and commonly form thin layers representing bedding or mineralized fracture cleavage (Pl. 3, figs. 3, 4). Some of the quartzite is very white, with practically no dark minerals.

Quartz-mica schist is as abundant as quartzite. The microscopic texture is a granoblastic mosaic of quartz and plagioclase, with abundant, well-oriented muscovite and biotite flakes. Numerous porphyroblasts of biotite and garnet are commonly concentrated along the bedding, or in older fracture cleavage planes. In the western part of the formation

porphyroblasts average 1 to 2 millimeters in diameter, but become much smaller eastward.

Mica schist is less common. Muscovite is generally predominant over biotite, giving the schist a spotted, silvery appearance, and garnet is common.

*Piermont member.*—Near the village of Piermont, light-colored rocks essentially similar to the rest of the Albee are interbedded with medium- to dark-gray beds, which commonly carry staurolite, and resemble typical metamorphosed sediments in the Orfordville formation. Thin sections of the dark beds reveal the same kind of disseminated carbon that is characteristic of the Orfordville. The proportion of light-colored to dark-colored beds varies widely in different exposures, and there is some gradation between the two types, but the combination of light and dark beds is distinctive. As these rocks are overlain and underlain by rocks of typical Albee lithology, they must be considered as belonging to a member within the Albee formation.

The arrangement of minerals in the darker beds of the Piermont member is similar to that in the Orfordville formation. Muscovite occurs in tiny flakes parallel to the schistosity. Easily visible biotite porphyroblasts are abundant, and garnet somewhat less so. Staurolite crystals one quarter to 2 inches long are common in the darkest beds west and northwest of Peaked Mountain, and kyanite is present in the immediate neighborhood of Piermont village.

Beds of fine conglomerate are characteristic of the Piermont member. The more common type consists of abundant pebbles of granular quartz, half an inch in diameter, embedded in dark gray, quartzose schist. Locally there are rounded fragments of light-colored gneiss similar to metamorphosed pyroclastics in the Orfordville formation. Numerous conglomerate beds, from 1 inch to 2 feet thick, are interbedded in dark-gray schist, especially in the lower half of the member. Associated with this conglomerate are a few thin beds of coarse, pebbly quartzite containing, besides quartz pebbles, sparse, subangular fragments of fine-grained quartzite and dark schist. Several of these beds, as much as 5 feet thick, also mark the top of the Piermont member at several places between Bean Brook and the north boundary of the quadrangle.

*Thickness.*—Inasmuch as it is impossible to measure in the usual manner the thickness of beds as greatly deformed as those in the Albee formation, crude estimates must suffice. These are based upon measurements of the breadth of outcrop along the axis of a syncline extending from the Mt. Cube quadrangle into the Moosilauke quadrangle, in which

FIGURE 1. BIOTITE-QUARTZ DIORITE
Of Mascoma group intruding hornblende-quartz diorite, marked by large hornblende crystals. Note indefinite contact cutting across specimen from upper left to lower center.

FIGURE 2. BIOTITE-QUARTZ DIORITE
Of Mascoma group and hornblendic inclusion Holts Ledge.

FIGURE 3. BEDDING AND RELIC FRACTURE CLEAVAGE
Preserved by crystallization of biotite in the fractures. Albee formation, 1 mile northwest of Peaked Mountain.

FIGURE 4. FOLDED RELIC FRACTURE CLEAVAGE IN ALBEE FORMATION
Mineralized fractures have been curved by shear parallel to the bedding visible in the specimen. East slope Sunday Mountain.

MASCOMA GROUP AND RELIC FRACTURE CLEAVAGE

is exposed a nearly complete section, from the lowest to the highest formations in western New Hampshire. Using as a basis for estimate the 2000 feet assigned by Billings (1937, p. 480) to the Ammonoosuc formation, the thickness of the Albee formation is roughly 5000 feet. The lower 2000 feet of the Albee is exposed in the Mt. Cube quadrangle. This includes the Piermont member, which is about 500 feet thick southwest of Peaked Mountain, increasing westward to perhaps 1000 feet at the Ammonoosuc thrust, and pinching out 3 miles to the east.

*Age.*—The Albee formation is definitely pre-middle Silurian (Billings, 1937, p. 475), and as it overlies the presumably middle Ordovician Orfordville formation, it is tentatively assigned to the upper Ordovician.

#### AMMONOOSUC VOLCANICS

*Correlation.*—The upper part of the Albee formation and part of the sequence above it are cut out in the Mt. Cube area by the Northey Hill thrust. East of the thrust, the Ammonoosuc volcanics of Billings (1937, p. 475-480) constitute the lowest stratigraphic unit, the lower part of which is displaced by the Oliverian intrusives. Thus in the Mt. Cube area it is estimated that the upper 3000 feet of the Albee formation and the lower half of the Ammonoosuc are missing. The remaining part of the Ammonoosuc volcanics consists of fine-grained biotite gneiss, hornblende gneiss, and amphibolite.

*Lithology and petrography.*—In the field the various lithologic types are distributed irregularly and apparently occur in large lenses, a hundred or more feet thick, and many hundreds of feet long. Bedding is rarely seen, although intercalations of biotite gneiss and hornblende gneiss or amphibolite, several inches to 10 or 20 feet thick, occur. In one or two places are outcrops which might represent conglomerates.

Fine-grained biotite gneiss is most abundant, comprising about 70 per cent of the formation. (*See* modes in Table 4.) It represents chiefly quartz latite and dacite, generally with excess quartz from detrital sources. The gneiss is mostly light-colored, with sparsely scattered biotite, and commonly contains very small phenocrysts of plagioclase. Locally, darker varieties richer in biotite are present.

Amphibolite, representing metamorphosed basalt, amounts to about 30 per cent of the formation. It is uniformly dark green, contains abundant tiny needles of hornblende, and has a variable texture. In much of the amphibolite near the contact of the Oliverian intrusives, epidote is abundant. Locally the amphibolite is streaked with layers deficient in hornblende, and the term hornblende gneiss is appropriate.

*Thickness.*—The maximum thickness of the Ammonoosuc volcanics in the Mt. Cube area, 1000 feet, is found along the east side of Smarts Mountain. Where the breadth of outcrop is less, the formation has been

TABLE 4.—*Modes of the Ammonoosuc volcanics*

| | 1 | 2 | | 3 | 4 | 5 |
|---|---|---|---|---|---|---|
| Quartz | 29 | 40 | 25–50 | 30 | 3 | 5 |
| Plagioclase | 40 | 56 | 40–69 | 52 | 57 | 28 |
| Potash feldspar | 20 | tr | 0–tr | | | |
| Muscovite | tr | tr | 0–1 | | | |
| Biotite | 10 | 1 | 0–5 | 2 | tr | tr |
| Chlorite | tr | 3 | 0–9 | | 8 | 2 |
| Garnet | | tr | 0–tr | | | |
| Amphibole | | tr | 0–1 | 16 | 30 | 64 |
| Epidote | 1 | tr | 0–tr | | | |
| Apatite | tr | tr | 0–tr | tr | tr | tr |
| Sphene | tr | tr | 0–tr | tr | | |
| Magnetite | | tr | 0–1 | tr | 2 | 1 |
| Totals | 100 | 100 | | 100 | 100 | 100 |
| Character of plagioclase Ab | 65 | 77 | | 80 | 63 | labradorite |
| Character of plagioclase An | 35 | 23 | | 20 | 37 | |
| Grain size, mm. Groundmass | 0.15 | 0.15 | | 0.2 | 0.1 | 0.2 |
| Grain size, mm. Porphyroblasts * | 2.0 | 2.0 | | 0.4 | 0.4 | |

* Includes crystals in crystal tuff.

1. Biotite gneiss (quartz latite), average of 3 sections.
2. Biotite gneiss (dacite), average and range of 5 sections.
3. Amphibole gneiss (dacite), average of 2 sections.
4. Amphibole gneiss (andesite), average of 2 sections.
5. Amphibolite (basalt), average of 3 sections.

either thinned tectonically, or cut out by intrusives. The total thickness of the Ammonoosuc volcanics in the Littleton-Moosilauke area is 2000 feet (Billings, 1937, p. 480).

*Age.*—The Ammonoosuc formation directly underlies the Clough formation of lower or middle Silurian age. It is separated from the Clough in the Littleton-Moosilauke area by a considerable unconformity, and its age is considered to be upper Ordovician.

### CLOUGH FORMATION

*Correlation.*—The name Clough conglomerate was given by Billings (1937, p. 481) to interbedded quartzite and quartz conglomerate, 150 feet thick, which overlies the Ammonoosuc volcanics in the Littleton-Moosilauke area. This formation was then recognized as an extremely useful key to the stratigraphy and structure in western New Hampshire, and it has since been traced southward nearly to the Massachusetts State line, a distance of about 100 miles. In the Mt. Cube quadrangle, the conglomerate beds are subordinate, and considerable quartz-mica schist is included; hence the term Clough formation has been adopted.

*Lithology.*—The Clough formation in the Mt. Cube area contains quartzite, quartz conglomerate, and quartz-mica schist, with many gradations between these types.

Quartzite is the most conspicuous type; massive, white or bluish-gray outcrops are characteristic of the lower part of the formation. In the upper part it is interbedded with quartz-mica schist (Pl. 1, fig. 1). It is composed of coarse-grained, vitreous quartz, with a small amount of muscovite, and traces of garnet, kyanite, and chlorite.

Quartz conglomerate is interbedded with quartzite largely in the lower beds of the formation. It contains abundant pebbles of coarse-grained, glassy quartz, generally squeezed into elliptical or cigar-shaped bodies 1 to 3 inches long. On the north end of Holts Ledge the pebbles have been so flattened in the plane of schistosity that they appear like elliptical pancakes, of which the longest axis is 8 or 10 times the shortest. The matrix is locally quartz-muscovite schist in which the pebbles are readily seen; but generally both pebbles and matrix consist almost exclusively of quartz so recrystallized that it is difficult to distinguish between them.

In the upper part of the Clough formation on Mt. Cube, Lambert Ridge, and Smarts Mountain, quartz-mica schist is abundant. It is rather coarse-grained, with well-developed schistosity and is commonly interbedded with quartzite layers a foot or so thick. Minerals present besides quartz are: muscovite (10 to 15 per cent), biotite (1 to 10 per cent), and euhedral garnets, 2 to 6 millimeters in diameter. On Mt. Cube and Piermont Mountain staurolite and kyanite occur in the schist.

On the southeast slope of Smarts Mountain, where the Clough formation is unusually thick, the upper part consists of quartz-mica schist interbedded with buff and gray quartzites resembling those in the overlying Fitch formation. Lime-silicate minerals are present locally, but no truly calcareous beds were found.

*Thickness.*—The present thickness of the Clough formation varies greatly in the Mt. Cube area, and, although its original thickness is hard

to estimate because of the possibility of folding, squeezing, and repetition by thrusting, it is certainly considerably greater than in the Littleton-Moosilauke area. The thickness of the most reliable sections in the Mt. Cube area are given in Table 5.

TABLE 5.—*Thickness of Clough formation in Mt. Cube area*

| Locality | Average dip | Approximate thickness | Remarks |
|---|---|---|---|
| Holts Ledge..... | 24°NW. | 400 feet | Little folding; tectonically thinned at least 35 percent* |
| Lambert Ridge..... | 40°W. | 600 feet | Considerable open folds; less squeezing; minimum. |
| East side Smarts Mtn.. | 32°E. | 1200 feet | Little folding seen; may be repeated by thrusts; base of Fitch cut out. |

* Based on calculations from squeezed pebbles. (*See* text.)

The considerable increase in the thickness of the Clough formation in the Mt. Cube area compared to the Littleton-Moosilauke area is believed to have occurred at the expense of the overlying Fitch formation for two reasons: (1) in the upper part of the Clough formation on Smarts Mountain the presence of lithologic types resembling those in the Fitch, and containing lime-silicate minerals; (2) a corresponding decrease in the thickness of the Fitch where the Clough is thickest. This cannot be shown on Smarts Mountain for the Fitch is cut out by the Bethlehem gneiss, but it is evident in the Mascoma quadrangle, where, according to Chapman (1939, p. 139) the Clough has a maximum thickness of 1200 feet and the Fitch, a thickness of less than 100 feet.

*Age.*—The Clough formation in the Littleton-Moosilauke area underlies fossiliferous middle Silurian strata, and unconformably overlies rocks of probable Ordovician age (Billings, 1937, p. 483). Hence, its age is either middle or lower Silurian. The interfingering relationship between the Clough and the Fitch formations mentioned above suggests that certainly the upper part of the Clough, and perhaps all of it, belongs to the middle rather than the early Silurian.

### FITCH FORMATION

*Correlation.*—Immediately overlying the Clough formation in the Mt. Cube quadrangle is a series of calcareous quartz-mica schists and sandstones, with some impure marble. These have the same lithologic char-

acteristics and stratigraphic position as the Fitch formation of the Littleton-Moosilauke area (Billings, 1937, p. 483-486).

*Lithology.*—The following types are recognized in the Fitch formation in the Mt. Cube quadrangle:

(1) Quartz-mica schist, commonly calcareous
(2) Calcareous sandstone
(3) Impure marble
(4) Quartzite and quartz conglomerate
(5) Lime-silicate rocks, including:
    biotite-actinolite schist
    epidote-diopside-garnet granulite
    diopside-feldspar granulite
    actinolite-feldspar granulite
    tremolite-quartz-feldspar granulite

Modes of these rocks are given in Table 6.

Quartz-mica schist is the most abundant type. It is brownish gray, highly schistose, and locally contains numerous pods of buff marble. Quartz, much reddish-brown biotite, and some muscovite make up the bulk of the rock. Disseminated calcite is commonly present but not conspicuous.

Calcareous sandstone forms beds 5 feet or less thick, interbedded with quartz-mica schist. It is light gray, somewhat schistose when fresh, buff colored and rather friable when weathered. Brown biotite and muscovite may be abundant, the latter commonly forming porphyroblasts. Calcite ranges from very little to as much as 30 per cent.

Marble is coarse-grained, bluish-gray when fresh, buff when weathered. Locally beds as much as 4 feet thick are found, but these comprise a very small part of the formation. Porphyroblastic muscovite is commonly present,—also disseminated biotite and quartz. Pure, white quartzite, locally carrying small quartz pebbles, is found in the Fitch at several places, but it is nowhere more than a few tens of feet thick.

Throughout the quadrangle greenish rocks containing abundant actinolite or diopside are found in beds a few feet thick at the very top of the Fitch, and at two places considerably greater amounts occur. They are rare types in the formation as a whole, are all characterized by various minerals relatively rich in magnesia, and probably represent original dolomitic sediments.

*Thickness.*—The limits of the Fitch formation in the Mt. Cube area are considered to coincide with the extent of calcic beds overlying the

Clough formation. Well-exposed sections are scarce, but the best data indicate a thickness between 400 and 600 feet for most of the area. Data from the best sections are given in Table 7.

TABLE 6.—*Modes of the Fitch Formation*

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| Quartz | 5 | 67 | 20 | tr | 8 | 6 |
| Calcite | 93 | 25 | | 10 | tr | 3 |
| Muscovite | 1 | 3 | | | tr | |
| Biotite | 1 | 5 | 12 | | | tr |
| Plagioclase | | | 30 | | 12 | 33 |
| Potash feldspar | | | | | 34 | 20 |
| Garnet | | | | 25 | | |
| Actinolite | | | 18 | 3 | tr | 33 |
| Hornblende | | | | | | |
| Diopside | | | 3 | 42 | 44 | tr |
| Epidote and clinozoisite | | tr | 17 | 20 | 1 | 5 |
| Apatite | | | | | | |
| Sphene | | | tr | tr | 1 | tr |
| Total | 100 | 100 | 100 | 100 | 100 | 100 |
| Character of plagioclase Ab | | | alb. | 65 | 45 | 50 |
| Character of plagioclase An | | | | 35 | 55 | 50 |
| Grain size mm. | 0.4 | 0.3 | 0.05 | 0.1 | 5.0 | 0.1 |

1. Arenaceous marble, 1 section.
2. Calcareous sandstone, average of 2 sections.
3. Biotite-actinolite schist, 1 section.
4. Epidote-diopside-garnet granulite, 1 section.
5. Diopside-feldspar granulite, 1 section.
6. Actinolite-feldspar granulite, average of 2 sections.

*Age.*—No fossils have been found in the Fitch formation in the Mt. Cube area, but marine fossils in the Littleton-Moosilauke area indicate it is Niagaran (Billings and Cleaves, 1934, p. 415-418).

### LITTLETON FORMATION

*Correlation.*—Schists which overlie the Fitch formation in the Mt. Cube quadrangle can be traced northeastward, through the Woodsville quadrangle into the Moosilauke quadrangle, where they join with rocks

assigned to the Littleton formation by Billings (1937, p. 487-495). A large part of this belt in the Mt. Cube area is cut out by the Bethlehem gneiss.

*Lithology.*—The rocks of the Littleton formation are rather uniformly gray or greenish gray, highly schistose, and generally fine-grained. The

TABLE 7.—*Thickness of Fitch formation in Mt. Cube area*

| Location | Average dip | Thickness (in feet) | Remarks |
|---|---|---|---|
| Stream flowing SW. from Piermont Mtn. | 50°W. | 300 | Good control on top and bottom, but folded and thinned tectonically. |
| North slope Winslow Ledge | 26°N. | 400 minimum<br>600 maximum | Good control at base, poor at top. Error due to open folds would reduce figure slightly. |
| North end Holts Ledge, at road | 22°NW. | 200 maximum | Poor control, only 50 feet exposed. Probable thinning, 50 per cent. |
| North end Moose Mtn. | 30°W. | 70 maximum | Little folding, but tectonic thinning may be important. |

metamorphic texture becomes markedly coarser, however, in the eastern part of the belt. Stratification is generally not visible, though very regular quartzose beds, a few inches thick are not uncommon. The rocks are largely classed as quartz-mica schist and mica schist, though some schistose quartzite occurs. In the lower part of the formation, just above the Fitch, numerous, very thin, pink beds composed of fine-grained garnet, quartz and apatite, apparently represent metamorphosed, impure phosphatic beds.

Quartz-mica schist and mica schist are composed largely of quartz, biotite, muscovite, and chlorite; in some specimens chlorite is very abundant and no biotite is present (Table 8). These rocks are very similar megascopically to schists in the Orfordville formation, but may be distinguished from them by large amounts of chlorite and by the absence of the disseminated carbon so characteristic of the Orfordville sediments. The small red or pink garnets, present in most outcrops, are commonly altered to chlorite, especially in the western part of the belt. East of the Indian Pond pluton of Bethlehem gneiss, the schist becomes considerably coarser, and sporadic staurolite crystals as much as half an inch long appear.

Quartzite, containing more than 80 per cent quartz, is common but subordinate in amount. It contains considerable mica similar to that in the schists and is usually cleavable. Locally, where the cleavage is regular, the quartzite has been quarried for whetstones.

TABLE 8.—*Modes of the Littleton formation*

| | | 1 | | 2 |
|---|---|---|---|---|
| Quartz | | 60 | 55–70 | 40 |
| Muscovite | | 18 | 15–20 | |
| Biotite | | 12 | 5–20 | |
| Chlorite | | 10 | 0–25 | |
| Garnet | | tr | 0– 1 | 55 |
| Apatite | | tr | 0–tr | 5 |
| Tourmaline | | tr | 0–tr | |
| Magnetite | | tr | tr | |
| Totals | | 100 | | 100 |
| Grain size, mm. | Groundmass | 0.1 | | 0.02–0.1 |
| | Porphyroblasts | 0.8 | | |

1. Quartz-mica schist, average and range of 3 sections.
2. Pink garnet layers, average of 2 sections.

*Thickness.*—The upper part of the Littleton formation in the Mt. Cube area is cut out along the Northey Hill thrust. Moreover, because of the intense folding, the actual thickness of beds exposed can only be guessed. Of the 5000 feet estimated for the formation in the Littleton-Moosilauke area (Billings, 1937, p. 493) probably the lower 2000 or 3000 feet are exposed in the Mt. Cube area.

*Age.*—Fossils found 2500 to 3000 feet above the base of the Littleton formation in the Littleton-Moosilauke area belong to the Oriskany epoch of the Devonian (Billings and Cleaves, 1934, p. 418-422).

## PLUTONIC AND HYPABYSSAL IGNEOUS ROCKS

### GENERAL STATEMENT

Plutonic rocks occupy about 30 per cent of the Mt. Cube area. They are chiefly found in large, gneissic bodies, intruded during the earlier and later stages of the main orogeny, although one older body lies in the

northwest part of the area. Granodiorite and quartz monzonite [1] are by far the most abundant petrographic types, associated with small amounts of granite and quartz diorite. In accordance with the classification of

TABLE 9.—*Estimated modes of the Fairlee quartz monzonite*

| | 1 | 2 | | |
|---|---|---|---|---|
| Quartz | 23 | 29 | 27 to | 32 |
| Potash feldspar | 50 | 29 | 23 | 34 |
| Plagioclase | 13 | 19 | 17 | 20 |
| Zoisite | 3 | 5 | 3 | 8 |
| Sericite | 4 | 7 | 6 | 8 |
| Muscovite | tr | 1 | 0 | 2 |
| Biotite and chlorite | 4 | 10 | 8 | 14 |
| Apatite | tr | tr | | |
| Sphene | | tr | | |
| Magnetite | | tr | | |
| Rutile | | tr | | |
| Zircon | tr | tr | | |
| Tourmaline | 3 | | | |
| Calcite | tr | tr | | |
| Leucoxene | tr | tr | | |
| Totals | 100 | 100 | | |
| Character of plagioclase | $Ab_{95}An_5$ | $Ab_{95}An_5$ | | |
| Probable original plagioclase | $Ab_{77}An_{23}$ | $Ab_{75}An_{25}$ | | |

1. Granite in Fairlee quartz monzonite body, 1 section.
2. Fairlee quartz monzonite, average and range of 4 sections.

Billings (1937, p. 469-470) and others, these plutonic rocks are grouped into three comagmatic series: the oldest, the Highlandcroft magma series, probably late Ordovician; the Oliverian magma series, intruded during the earlier stages of the late Devonian orogeny; and the New Hampshire

[1] In the classification of the plutonic rocks, Johannsen's granite and granodiorite families have been modified according to the following proportions of potash feldspar to total feldspar: granite, 95 to 67 percent; quartz monzonite, 66 to 33 percent; granodiorite 32 to 5 percent; quartz diorite, less than 5 percent. Muscovite has been added directly to potash feldspar, on the ground that original potash commonly appears in muscovite resulting from deuteric crystallization or metamorphism.

magma series, intruded during the later stages of the orogeny. Of the hypabyssal rocks, a group of metamorphosed basaltic sills preceded the New Hampshire magma series; and a later group of unmetamorphosed gabbros and related rocks may belong to a fourth comagmatic group, the White Mountain magma series (Chapman and Williams, 1935, p. 504). From evidence in the Mt. Cube area, the writer is inclined to believe that the Oliverian magma series and the New Hampshire magma series really belong to one comagmatic group.

### FAIRLEE QUARTZ MONZONITE

One body, about 0.7 mile wide and 6.5 miles long, is correlated on the basis of lithology with the Highlandcroft magma series in the Littleton-Moosilauke area. This body is exposed prominently on cliffs at Fairlee, Vermont in the central part of the quadrangle, where it was studied by C. H. Richardson and called the Fairlee granite gneiss (Foyles, 1928, chart opp. page 288; also personal communication). It is a coarse-grained, granitoid rock, generally greenish gray, with local pink tinges. At several places it is subporphyritic. A weak to moderate foliation is common, and locally at the margin of the body the rock is highly schistose.

The essential minerals of the quartz monzonite are bluish-gray quartz, perthitic microcline with Carlsbad twinning, saussuritized plagioclase, chlorite and sericite (Table 9). Quartz grains are commonly fractured, with pronounced strain shadows and sutured boundaries; moreover, the microcline is broken, and locally much altered to sericite. Remnants of green biotite contain abundant sagenitic sphene and are considerably altered to bright-green penninite. Assuming that the zoisite in the saussuritized plagioclase represents anorthite, and that there has been no gain or loss of lime during metamorphism under low-grade conditions, the original plagioclase ranged from oligoclase ($Ab_{83}An_{17}$) to andesine ($Ab_{66}An_{34}$), and the original rock was a quartz monzonite.

The Fairlee quartz monzonite intrudes the Albee formation of Ordovician age, although the relations are obscured by intense deformation along the contact. It has suffered pronounced dynamothermal metamorphism, similar to that which affects the Albee and other rocks around it. In the Littleton-Moosilauke area, Billings (1937, p. 518; also Pl. 1) has shown that the Highlandcroft granodiorite was exposed on the erosion surface immediately below the middle Silurian Fitch formation. The Fairlee quartz monzonite is therefore believed to have been intruded during the late Ordovician.

### OLIVERIAN MAGMA SERIES

*General statement.*—The term, Oliverian magma series, was first used by Billings (1935, p. 26) to designate a comagmatic group of plutonic

rocks. The series, intrusive into the rocks immediately above it, occupies the center of domical uplifts and is represented in the Mt. Cube area by the Baker Pond gneiss, Smarts Mountain granodiorite, and the Mascoma group, which consists of quartz diorite, granodiorite, quartz monzonite, and granite.

*Baker Pond gneiss.*—This rock is a portion of a larger body, exposed in the Rumney quadrangle and named by Page (1937). Most of the Baker Pond gneiss is a moderately foliated gray biotite-quartz monzonite (Table 10, column 1). Southwest of Upper Baker Pond it contains 10 per cent of white or pinkish microcline phenocrysts, 3 to 12 millimeters long. In the Ammonoosuc formation near the border of the body are sills of finer grained, more foliated gneiss, some of which are quartz monzonite, and some granodiorite with microcline phenocrysts 1 or 2 millimeters long. A border phase of dark, greenish quartz diorite gneiss occupies a zone 1400 feet wide at the south border of the body. It intrudes the Ammonoosuc formation, but its relations to the biotite-quartz monzonite are unknown.

*Smarts Mountain granodiorite.*—This rock consists almost entirely of light-gray, fine- to medium-grained biotite granodiorite, with a distinctly sugary texture. Near the borders it is somewhat finer textured and locally carries numerous lenticular inclusions of biotite-hornblende schist derived from the Ammonoosuc volcanics.

Data from thin sections are given in Table 10. Quartz and untwinned oligoclase make up nearly 85 per cent of the average granodiorite, forming a markedly granoblastic mosaic. About 10 per cent of microcline is generally present, and tends to be interstitial. The biotite forms more or less parallel, subcircular plates less than 1 millimeter across; it is pleochroic; X and Y = light greenish yellow; Z = dark olive green. In the analysed rock (Table 10, column 5) the gamma index of biotite is 1.647 and the birefringence is .043; these data indicate a variety rich in alumina with the FeO : MgO ratio about 3:1. Magnetite octahedra, as much as 2 millimeters across, are sparsely scattered in nearly every outcrop.

*Mascoma group.*—The rocks of the Mascoma group appear in the northern part of a great dome, 5 miles by 17 miles, most of which lies in the Mascoma quadrangle. In the Mt. Cube quadrangle, these are all rather gneissic, granular rocks, generally with a characteristic olive-green biotite as the chief dark mineral. On the basis of field and laboratory study, they have been divided among four types: (1) granite, (2) quartz monzonite, (3) sodic granodiorite, and (4) quartz diorite. In the

TABLE 10.—*Modes of the Oliverian magma series*

| | | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 | 14 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Quartz | | 25 22–32 | 26 | 5 | 35 25–40 | 36.1 | 40 33–47 | 30 30–30 | 48 40–55 | 42.5 | 39 | 29 22–33 | 51 | 25 | 32 20–44 |
| Potash feldspar | | 30 27–32 | 15 | 7 | 11 6–15 | 6.7 | 42 40–45 | 24 24–25 | 7 5– 9 | 5.0 | 11 | | | tr | tr 0–tr |
| Plagioclase | | 30 25–40 | 45 | 54 | 48 40–62 | 51.4 | 15 5–25 | 33 32–33 | 41 35–47 | 45.6 | 42 | 64 63–64 | 42 | 53 | 57 50–67 |
| Muscovite | | tr 0–tr | 1 | | tr 0–tr | | tr tr | tr 0–tr | tr tr– 1 | | tr | tr 0–tr | 1 | | |
| Biotite | | 11 7–15 | 12 | 10 | 5 2–06 | 5.6 | 3 2– 3 | 11 10–12 | 4 3– 4 | 6.2 | 8 | 6 3–10 | 6 | 10 | tr 0– 1 |
| Chlorite | | | | | | | | | | | | tr 0– 2 | | | tr tr |
| Amphibole | | | | 20 | | | | | | | | | | 12 | 10 6–12 |
| Epidote | | 4 2– 5 | 1 | 4 | tr 0–tr | | | 2 1– 3 | tr tr | | | 1 0– 3 | | | 1 0– 2 |
| Magnetite | | tr 0–tr | | tr | 1 tr–2 | 0.9 | tr tr | tr 0–tr | tr 0–tr | 0.5 | | tr 0–tr | tr | tr | tr 0–tr |
| Apatite | | tr tr–tr | tr | tr | tr tr | tr | tr tr | tr 0–tr | tr tr | 0.1 | tr | tr tr | tr | | tr tr |
| Sphene | | tr tr–01 | tr | tr | tr tr | 0.2 | tr 0–tr | tr tr | tr tr | tr | | tr tr | | | tr 0– 1 |
| Zircon | | | | | | | | | | | | tr 0–tr | tr | | |
| Garnet | | | | | | | | | | | | | tr | | tr 0–tr |
| Scapolite | | | | tr | | | | | | | | | | | |
| Pyrite | | tr 0–tr | tr | | | | | | | | | | | | |
| Calcite | | | | | | | | | | 0.1 | | | | | |
| Totals | | 100 | 100 | 100 | 100 | 100.9 | 100 | 100 | 100 | 100.0 | 100 | 100 | 100 | 100 | 100 |
| Character of plagioclase | Ab | 65 60–70 | 63 | 68 | 80 88–75 | 76 | 68 60–75 | 63 60–65 | 81 78–84 | 80 | 65 | 68 60–85 | 79 | 80 | 70– 59–80 |
| | An | 35 40–30 | 37 | 32 | 20 12–25 | 24 | 32 40–25 | 37 40–35 | 19 22–16 | 20 | 35 | 32 40–15 | 21 | 20 | 30 41–20 |
| Average grain size in mm | | 0.4 0.2–0.5 | 0.3 | 0.6 | 0.6 0.4–1.0 | | 0.5 0.2–1.0 | 0.4 0.3–0.6 | 0.6 0.2–2.0 | | 0.4 | 0.2–0.4 | 0.5 | 0.3 | 0.2–0.4 |

1. Baker Pond gneiss: quartz monzonite, average and range of 3 sections.
2. Baker Pond gneiss: granodiorite, 1 section.
3. Baker Pond gneiss: mixed type, 1 section.
4. Smarts Mountain granodiorite: average and range of 6 sections.
5. Smarts Mountain granodiorite: mode calculated from chemical analysis in Table 11, column 2.
6. Mascoma group: granite, average and range of 2 sections.
7. Mascoma group: quartz monzonite, average and range of 2 sections.
8. Mascoma group: oligoclase granodiorite, average and range of 2 sections.
9. Mascoma group: oligoclase granodiorite, mode calculated from the chemical analysis in Table 11, column 3.
10. Mascoma group, quartz diorite zone: granodiorite 1 section.
11. Mascoma group, quartz diorite zone: biotite-quartz diorite, average and range of 3 sections.
12. Mascoma group, quartz diorite zone: siliceous biotite-quartz diorite, average of 2 sections.
13. Mascoma group, quartz diorite zone: biotite-quartz diorite with hornblende, 1 section.
14. Mascoma group, quartz diorite zone: hornblende-quartz diorite, average and range of 3 sections.

field these types grade into one another, so that definite boundaries cannot be drawn between them. Yet a distinct areal distribution can be shown on the map. Data from thin sections are available in Table 10, columns 6 to 8 and 10 to 12.

Granite occurs in a small, nearly circular area in the western part of the main body. It is light gray, or pinkish, and typically porphyritic, resulting in a coarse-grained appearance. The groundmass is medium-grained, distinctly granoblastic, and is composed of granular quartz aggregates in a mosaic of quartz, oligoclase (locally myrmekitic), and interstitial microcline. Anhedral, ragged crystals of microcline, 10 to 20 millimeters long, are abundant. A relatively small amount of olive-green biotite is the only significant dark mineral. A moderate foliation is produced by parallel arrangement of biotite flakes and flattened quartz aggregates.

Quartz monzonite occupies the central part of the main mass in the Mascoma and Mt. Cube quadrangles. It is medium-grained, gneissic, with considerably more olive-green biotite than the granite. Large crystals of microcline are common, but are smaller than in the granite, and amount to less than 1 per cent of the rock. The texture of the groundmass is similar to that of the granite, but the plagioclase is andesine with more sodic margins. The quartz shows strain shadows and granulation. Biotite occurs in spotty aggregates of small grains. The rock as a whole strongly resembles parts of the Bethlehem gneiss (Pl. 2, fig. 4).

Sodic granodiorite occupies a broad, arcuate belt around the north and east sides of the granite and quartz-monzonite. It is medium-grained, gray, strongly foliated, composed largely of a friable aggregate of granular quartz and oligoclase. Thin, parallel laminae, composed of aggregates of small biotite flakes, are distributed 1 to 3 millimeters apart throughout the granodiorite, giving it a considerable schistosity (Pl. 2, fig. 2). Quartz aggregates are much flattened parallel to the biotite laminae. Magnetite octahedra, as in the Smarts Mountain granodiorite, are sparsely scattered through most outcrops.

The granodiorite is generally uniform, but the outer part contains inclusions of feldspathic hornblende-biotite schist, and layers several feet thick of light-colored gneiss, probably representing parts of the Ammonoosuc volcanics.

Thin sections of the granodiorite show a granoblastic mosaic of quartz and oligoclase, with interstitial microcline. The olive-green biotite is closely similar to that in the Smarts Mountain granodiorite; in the analysed specimen (Table 10, column 9), the maximum refractive index is 1.652. The plagioclase is sodic oligoclase ranging from $Ab_{85}An_{15}$ to $Ab_{80}An_{20}$, without zoning or twinning.

8. Mascoma group: oligoclase granodiorite, average and range of 2 sections.
9. Mascoma group: oligoclase granodiorite, mode calculated from the chemical analysis.
10. Mascoma group, quartz diorite zone: hornblende-quartz diorite, average and range of 3 sections.

TABLE 11.—*Chemical Analyses and modes of the Oliverian magma series and similar rocks*

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 72.72 | 74.30 | 76.53 | 72.11 | 73.52 | 73.27 | 65.01 | 61.32 |
| $TiO_2$ | 0.19 | .18 | .23 | .17 | .51 | .10 | .57 | .23 |
| $Al_2O_3$ | 14.06 | 14.04 | 12.82 | 15.25 | 12.86 | 15.51 | 15.94 | 16.95 |
| $Fe_2O_3$ | 1.02 | .78 | .78 | .64 | 1.48 | .33 | 1.74 | 2.39 |
| FeO | 1.23 | 1.27 | 1.24 | .84 | 1.54 | 1.14 | 2.65 | 4.29 |
| MnO | .06 | tr | .05 | .02 | .08 | tr | .07 | .05 |
| MgO | .60 | .47 | .45 | .38 | .57 | .15 | 1.91 | 2.84 |
| CaO | 1.46 | 2.74 | 1.98 | 1.98 | 2.08 | 2.74 | 4.42 | 5.56 |
| $Na_2O$ | 3.35 | 4.49 | 4.28 | 5.43 | 4.36 | 4.79 | 3.70 | 2.62 |
| $K_2O$ | 4.95 | 1.53 | 1.30 | 2.04 | 1.41 | 1.66 | 2.75 | 2.20 |
| $H_2O+$ | .33 | .27 | .24 | .66 | 1.21 | .68 | 1.04 | 1.22 |
| $H_2O-$ | n.d. | n.d. | 0.04 | .03 | .35 | n.d. | | |
| $P_2O_5$ | .04 | .02 | .05 | .06 | .07 | tr | .20 | .33 |
| $CO_2$ | n.d. | n.d. | .06 | .22 | n.d. | n.d. | | |
| S | .02 | n.d. | n.d. | .06 | none | n.d. | | |
| BaO | n.d. | n.d. | n.d. | 0.03 | n.d. | n.d. | | |
| Totals | 100.03 | 100.09 | 100.05 | 99.92 | 100.04 | 100.37 | 100.00 | 100.00 |
| S.G. | 2.633 | 2.661 | 2.675 | 2.676 | n.d. | | | |
| Quartz | 30.0 | 36.1 | 42.5 | 31.0 | 33.0 | | | |
| K feldspar | 51.7 | 6.7 | 5.0 | tr | | | | |
| Oligoclase | 13.0 | 51.4 | 45.6 | 54.0 | 53.0 | | | |
| Muscovite | 0.0 | | | 4.0 | 7.0 | | | |
| Biotite | 4.4 | 5.6 | 6.2 | 5.0 | 3.0 | | | |
| Hornblende | | | | | | | | |
| Magnetite and ilmenite | 0.7 | .9 | .5 | .4 | 1.0 | | | |
| Apatite | | tr | .1 | tr | tr | | | |
| Sphene | 0.2 | .2 | tr | | tr | | | |
| Epidote | | | | tr | tr | | | |
| Calcite | | | .1 | | | | | |
| Totals | 100.0 | 100.9 | 100.0 | 94.4 | 97.0 | | | |
| Character of plagioclase Ab | 70 | 76 | 80 | olig. | olig. | | | |
| An | 30 | 24 | 20 | | | | | |

1. Owls Head granite from the Moosilauke quadrangle (Billings, 1937, p. 556) Analyst: F. A. Gonyer.
2. (MTC 231) Smarts Mountain granodiorite from elevation 1620 feet in the next stream east of Mousley Brook, NW. slope Smarts Mountain. Analyst: F. A. Gonyer.
3. (MTC 358) Oligoclase granodiorite of the Mascoma group from beside the road 0.2 mile southwest of 1233 corner, south-central ninth, Mt. Cube quadrangle. Analyst: A. Willman.
4. Trondhjemite from Skavlien near Austberg, Norway. Röer analyst. (Johannsen, 1932, p. 383 and 385).
5. "Oligoclase granite" Barnes Hill, North Stonington, Conn. Also contains 0.02 NiO, which makes total 100.06. Steiger analyst (Johannsen, 1932, p. 380).
6. Williamsburg granodiorite (Emerson, 1917, p. 254) Eakins, analyst.
7. Average granodiorite (Daly, R. A., 1933, p. 15).
8. Average tonalite (Daly, R. A., 1933, p. 15).

Chemical analyses (Table 11) show that the granodiorite in both Smarts Mountain and Mascoma plutons contains an excessive amount of silica, and an unusually large amount of soda in relation to potash. Comparison of the granodiorites of the Oliverian magma series with pub-

lished modes and analyses given by Daly (1933) and Johannsen (1932) indicates that: (1) These granodiorites are special types, far from the average composition of any major family of plutonic rocks, such as the average granodiorite or tonalite; (2) there are few chemically similar types among the rocks of the world as reported by Johannsen. The most similar to the Oliverian granodiorites are a Norwegian *trondhjemite* and two "granites" and a granodiorite from New England.

Quartz diorite is confined to a complex zone between the three other bodies and the surrounding Ammonoosuc volcanics. The zone, well exposed in the 400-foot cliffs on Holts Ledge, includes four types. Biotite-quartz diorite, the most abundant rock, locally has hornblende needles as much as half an inch long. Siliceous biotite-quartz diorite, which is not common, has no hornblende but contains an unusually large amount of quartz. Hornblende-quartz diorite, which contains no biotite, is prominent on Holts Ledge Cliffs, where it is injected by biotite-quartz diorite (Pl. 3, figs. 1, 2; Pl. 1, fig. 3). Granodiorite, similar to that in the main granodiorite body, but with finer texture, occurs in sills near the top of the zone on Winslow Ledge and Holts Ledge. Many of these rocks are well foliated, and all are distinctly granular.

The quartz diorite zone is decidedly heterogeneous, and an adequate study of all of the rocks in it has not been made. On the whole it appears to represent a zone of mixed rocks between the major intrusives of the Mascoma group and the Ammonoosuc volcanics. At its inner margin, the zone, represented chiefly by biotite-quartz diorite, merges imperceptibly with the granodiorite of the core. This boundary is very difficult to distinguish in the field and is drawn so as to place the innermost hornblende-bearing rocks in the quartz-diorite zone, for hornblende does not appear in the granodiorite. The outer part of the zone contains numerous inclusions of coarse-grained hornblende-quartz diorite (Pl. 1, fig. 3) and of feldspathic biotite-hornblende schist, both intruded by biotite-quartz diorite. All the inclusions are slablike, the largest 2 to 3 feet thick and 50 feet long, lying parallel to the foliation in the quartz diorite. At some places the inclusions have been so recrystallized that they appear only as indistinct hornblendic patches in the enclosing rock. At the outer margin of the quartz diorite zone, amphibolite of the Ammonoosuc volcanics appears, injected by granodiorite sills up to 30 feet thick. The outer boundary of the zone has been arbitrarily placed at the outermost of these sills. Also, it is commonly marked in both the Smarts Mountain and Mascoma bodies by a narrow zone of intensely green, epidote-rich amphibolite in the volcanics.

The relation between the granodiorite and the granite and quartz monzonite of the Mascoma group indicates that the granodiorite is older

TABLE 12.—*Modes of the New Hampshire magma series*

| | | 1 | | 2 | | 3 | 4 | 5 | 6 | | 7 | 8 | | 9 | 10 | 11 | 12 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Quartz | | 33 | 30–34 | 31 | 23–40 | 32 | 31 | 32.7 | 35 | 35–35 | 37 | 34 | 25–45 | 40 | 34 | 34 | 30 |
| Potash feldspar | | 29 | 26–32 | 20 | 12–23 | 10 | 21 | 7.8 | 7 | 3–12 | 29 | 21 | 17–25 | 2 | 21 | 12 | 15 |
| Plagioclase | | 19 | 18–20 | 27 | 22–37 | 35 | 26 | 31.5 | 38 | 36–39 | 14 | 30 | 15–23 | 54 | 30 | 48 | 43 |
| Muscovite | | 10 | 7–13 | 7 | 3–10 | 2 | 7 | 10.3 | 4 | 3– 6 | 11 | 6 | 0–15 | 1 | 7 | 2 | 3 |
| Biotite | | 9 | 6–12 | 15 | 12–18 | 21 | 15 | 16.7 | 16 | 14–18 | 9 | 9 | 7–12 | 3 | 8 | 4 | 5 |
| Hornblende | | | | | | | | | | | | tr | 0– 4 | | tr | | |
| Epidote | | | | | | | | | | | | tr | 0–tr | | tr | , | 3 |
| Allanite | | | | | | | | | tr | 0–tr | | | | | | tr | |
| Magnetite | | tr | 0–tr | tr | 0–tr | tr | tr | | tr | 0–tr | | tr | 0–tr | | tr | tr | |
| Apatite | | tr | tr–0.5 | tr | tr | tr | tr | 0.5 | tr | 0–tr | tr | tr | 0–tr | | tr | tr | |
| Sphene | | tr | 0–tr | tr | 0–tr | tr | tr | 0.4 | tr | 0–tr | tr | tr | 0– 1 | | tr | tr | 1 |
| Zircon | | tr | tr | tr | 0–tr | tr | tr | | tr | tr | tr | tr | 0–tr | | tr | tr | |
| Garnet | | tr | 0–tr | tr | 0–tr | | tr | | | | | tr | 0– 1 | | tr | | |
| Tourmaline | | tr | 0–1 | | | | tr | | | | | tr | 0–tr | | tr | | |
| Calcite | | | | | | | | 0.1 | | | | | | | | | |
| Totals | | 100 | | 100 | | 100 | 100 | 100 | 100 | | 100. | 100 | | 100 | 100 | 100 | 100 |
| Character of plagioclase | Ab | 78 | 80–75 | 72 | 75–63 | 62 | 70 | 73 | 63 | 57–70 | oligoclase | 79 | 70–87 | 80 | 80 | 72 | 70 |
| | An | 22 | 20–25 | 28 | 25–37 | 38 | 30 | 27 | 37 | 43–30 | | 21 | 30–13 | 20 | 20 | 28 | 30 |
| Average grain size in mm | | 0.6 | | 0.6 | | 0.7 | | | 3.0 | 2.0–4.0 | 0.2 | | 0.03–0.3 | 0.07 | | 1.0 | 1.5 |

1. Bethlehem gneiss, Indian Pond pluton: granite, average and range of 3 sections.
2. Bethlehem gneiss, Indian Pond pluton: quartz monzonite, average and range of 6 sections.
3. Bethlehem gneiss, Indian Pond pluton: granodiorite, average of 2 sections.
4. Bethlehem gneiss, Indian Pond pluton: average of whole body, 11 sections.
5. Bethlehem gneiss, Indian Pond pluton: mode calculated from chemical analysis, Table 11, column 9.
6. Bethlehem gneiss, Mt. Clough pluton: granodiorite, average and range of 3 sections.
7. Bethlehem gneiss—fine-grained phase: granite, average of 2 sections.
8. Bethlehem gneiss—fine-grained phase: quartz monzonite, average and range of 7 sections.
9. Bethlehem gneiss—fine-grained phase: granodiorite, 1 section.
10. Bethlehem gneiss—fine-grained phase: grand average of 10 sections.
11. Haverhill granodiorite, average of 2 sections.
12. Haverhill granodiorite from Black Hill, Woodsville quadrangle, the type locality. Data supplied by Marland Billings.

than the other two. In a fairly well exposed section south of the trail, half a mile south of the summit of Holts Ledge, a gradational contact can be seen. Passing from the granodiorite to the granite, small metacrysts of microcline appear, and become larger and more abundant, until they supplant quartz as the most conspicuous mineral in the rock. The amount of biotite also decreases, and the strong foliation of the granodiorite diminishes. Though much less well exposed elsewhere, similar relations between the granodiorite and quartz monzonite exist, except that the amount of biotite increases in the quartz monzonite. These facts, and the microscopical study of the rocks, support C. A. Chapman's (1939, p. 167-170) conclusion that the granite, at least, is the result of hydrothermal replacement of granodiorite or quartz monzonite by microcline. In the Mt. Cube area, it does not appear likely that the quartz monzonite, too, is derived by replacement from the granodiorite.

The full significance of the quartz diorite zone in the Mt. Cube area is not clear. Much of the rock certainly represents greatly recrystallized parts of the Ammonoosuc volcanics. Amphibolite has been converted in this process to biotite-hornblende-andesine schist, and much of what has been called hornblende-quartz diorite probably represents recrystallized volcanic rocks. Sills of granodiorite and of biotite-quartz diorite also form a considerable part of the transition zone, but no granite or quartz monzonite sills were found.

*Age of the Oliverian magma series.*—In the Mascoma quadrangle, the middle Silurian Clough formation is cut by granite of the Mascoma group (C. A. Chapman, 1939, map). Moreover the lower Devonian Littleton formation is everywhere affected by the doming. Though it is conceivable that the intrusion took place independently, previous to the doming, this seems unlikely. Hence the age of the Oliverian rocks is considered to be post-lower Devonian, probably late Devonian. More detailed relations of the intrusion to the tectonic history of the region are discussed in a later paragraph.

### NEW HAMPSHIRE MAGMA SERIES

*General statement.*—Two large bodies of the Bethlehem gneiss of the New Hampshire magma series appear in the area,—the Indian Pond pluton and the Mt. Clough pluton (Fig. 6). There are also two smaller, stocklike bodies, for which the name Haverhill granodiorite is proposed. In addition numerous sills of fine-grained gneiss are associated with the Indian Pond pluton, and pegmatite bodies are associated with the Mt. Clough pluton.

*Bethlehem gneiss.*—The term Bethlehem gneiss has been used by Billings (1937, p. 504) for a group of intrusives of rather variable composi-

analysis, Table 11, column 9.
12. Haverhill granodiorite, from Black Hill, Woodsville quadrangle, the type locality. Data supplied by Marland Billings.

tion, which are, however, texturally and structurally similar and occur in great lenses, many miles long and 1 to 3 miles across, elongated parallel to the regional structure. This is a much more restricted use of the term than that employed by C. H. Hitchcock (1877, vol. 2, p. 104-111).

The gneiss in the Indian Pond pluton occupies an area 14½ miles by 1½ miles, extending northeast from Bear Hill in the southeastern part of Lyme township. It is dominantly medium-grained, gray biotite gneiss, cropping out generally in low, sheeted ledges. Anhedral phenocrysts of white microcline, 3 to 20 millimeters long, are common in parts of the body and constitute, where most abundant, 10 per cent of the rock; their distribution in the Indian Pond pluton is shown in Figure 3. In general the phenocrysts are most abundant in the northwestern part of the body and least common in the southeastern part. The writer has no explanation for this asymmetric arrangement. The foliation varies from prominent to weak and locally is absent. It is shown by subparallel, wavy laminae of mica, which commonly wrap around elongated quartz aggregates and feldspar phenocrysts. At several places a pronounced linear arrangement characterizes the micas, rather than a laminar one. Generally the transition from strong to weak foliation is very gradual; locally abrupt variations resemble shear zones. Aplite dikes, a few inches thick, traverse the gneiss locally; otherwise the gneiss is homogeneous, and associated pegmatites are very rare. Joints are present, but usually not numerous. The contacts of the gneiss are everywhere sharp, and the foliation at the margins is commonly more intense.

Microscopical study reveals that the composition of the gneiss in the Indian Pond pluton ranges from granite to granodiorite and that most of it is quartz monzonite. (*See* modes in Table 12.) These types differ in texture very little, but a close correlation exists between the abundance of microcline phenocrysts and the total amount of potash feldspar. Probably the varieties grade into one another, for no definite boundaries could be made out in the field. Microscopic examination shows an inequigranular mosaic of quartz, plagioclase, and microcline. Quartz appears in elongated, granular aggregates. Plagioclase grains are anhedral to subhedral, commonly twinned. Microcline occurs both in the groundmass and as subhedral phenocrysts with embayed margins, and, commonly, contains inclusions of quartz, plagioclase, and myrmekite. Large muscovite and reddish-brown biotite flakes are poorly to moderately well oriented in the sections; considerable fine-grained muscovite has formed by release of potash in the larger plagioclase grains. Small grains of apatite, sphene, and magnetite are disseminated in most sections. One place in the granite at the southern tip of the pluton is replete with rosettes of black tourmaline nearly an inch in diameter.

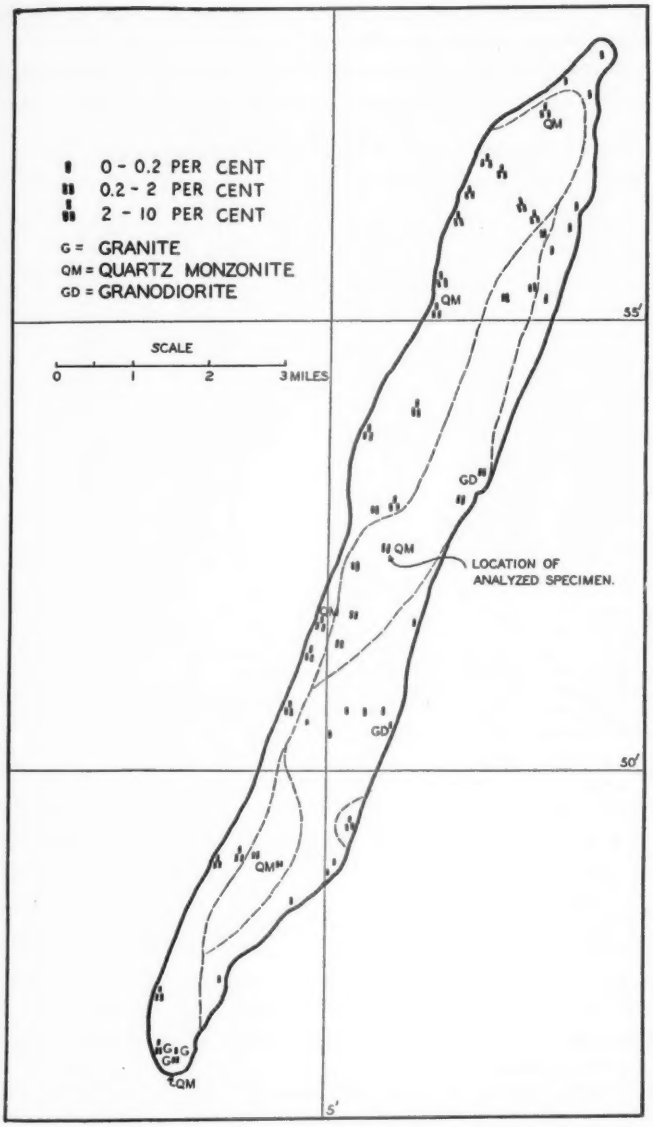


FIGURE 3.—*Distribution of phenocrysts in Indian Pond pluton*

Dashed lines indicate approximate boundaries only, to emphasize the asymmetrical distribution. Letter symbols refer to composition of hand specimens. Orientation of short, stubby lines has no significance.

A chemical analysis of the quartz monzonite gneiss of the Indian Pond pluton is given in Table 13, column 1.

Aplitic and pegmatitic differentiates from the Indian Pond pluton are rare, but small pods or lenses of both occur at two places in the neighbor-

TABLE 13.—*Chemical Analyses of Bethlehem Gneiss*

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 67.83 | 67.04 | 67.44 | 69.12 |
| $TiO_2$ | .71 | .57 | .67 | .28 |
| $Al_2O_3$ | 15.24 | 16.30 | 15.28 | 14.30 |
| $Fe_2O_3$ | .55 | .74 | 1.02 | 1.32 |
| FeO | 4.04 | 3.97 | 3.76 | 1.91 |
| MnO | .08 | .04 | .06 | .15 |
| MgO | 1.37 | 1.31 | 1.42 | 1.04 |
| CaO | 2.26 | 2.68 | 2.10 | 2.68 |
| $Na_2O$ | 2.67 | 2.83 | 2.76 | 2.49 |
| $K_2O$ | 3.96 | 4.03 | 4.37 | 5.44 |
| $H_2O+$ | .81 | .65 | .85 | .53 |
| $H_2O-$ | .02 | n.d. | n.d. | n.d. |
| $CO_2$ | .06 | none | none | .31 |
| $P_2O_5$ | .25 | .04 | .06 | .12 |
| S | n.d. | .11 | .06 | .06 |
| BaO | n.d. | none | n.d. | .18 |
| $ZrO_2$ | n.d. | n.d. | n.d. | none |
| Totals | 99.85 | 100.31 | 99.85 | 99.93 |
| S.G. | 2.748 | 2.717 | 2.681 | 2.683 |

1. (MtC 442) Quartz monzonite of Bethlehem gneiss, Indian Pond pluton, Quinttown-Orfordville road 0.7 mile northwest of Quinttown corner, Mt. Cube quadrangle. Analyst: A. Willman.
2. Bethlehem gneiss, village of Easton, Moosilauke quadrangle.
3. Bethlehem gneiss, 1.65 miles S. 38° E. of Swiftwater, Moosilauke quadrangle.
4. Bethlehem gneiss, 0.2 mile S. 42° W. of Gilmanton School, Littleton quadrangle.
Analyses 2, 3 and 4 from Billings (1937, p. 556).

hood of the body. A unique aplite occurs in two bodies, the largest 80 feet by 20 feet, just east of the road corner, 0.9 mile southwest of Bear Hill. It is composed of medium-grained, somewhat crushed quartz and albite (about 40 per cent and 59 per cent respectively) and muscovite (1 per cent).

The Bethlehem gneiss along the eastern boundary of the Mt. Cube quadrangle is petrographically similar to that in the Indian Pond pluton. However, it is more intensely foliated, is nonporphyritic, and is largely granodiorite, even approaching quartz diorite in composition.

Fine-grained, strongly foliated biotite gneiss forms narrow sills in the schist not far from the Indian Pond pluton. Most of the sills range in width from a few tens of feet to 100 feet, although one is 700 to 1200 feet wide and 3 miles long. They are shown individually on the map (Pl. 4), except for two regions, where they are especially numerous and are rep-

resented by a red overprint on the pattern for the Littleton formation.

Laboratory study shows that the sills range from granite to granodiorite, and that most of them are quartz monzonite; in other words,

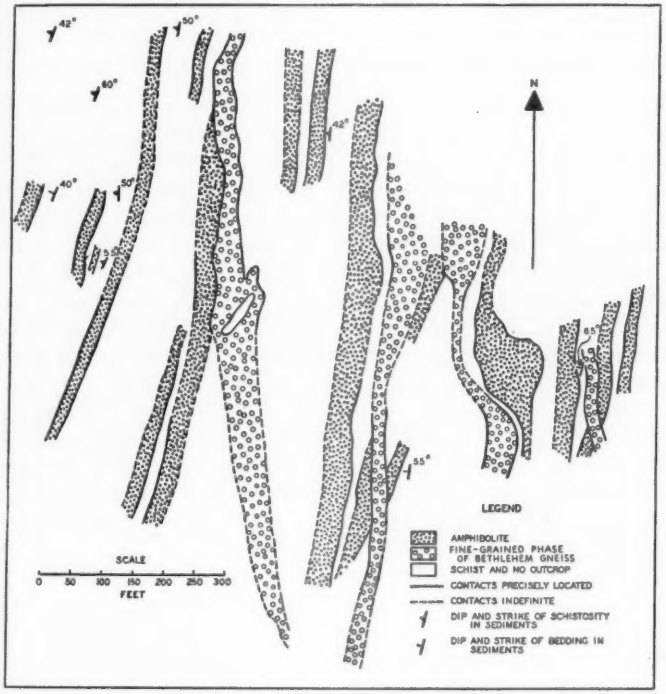


FIGURE 4.—*Sills and dikes of amphibolite and fine-grained phase of Bethlehem gneiss*

Intrude schist of Littleton formation, one-third mile northeast of Bench Mark 893, northeast corner of the Mt. Cube quadrangle. Primary purpose of mapping was to show that the fine-grained phase of the Bethlehem gneiss cuts the amphibolite; unfortunately no distinction was made between areas of schist and areas of glacial drift.

there is a distinct parallel between the composition of the sills and that of the Indian Pond pluton (Table 12).

In the extreme northeast corner of the quadrangle, the sills are well exposed, cutting at a slight angle the bedding of the Littleton formation as well as earlier amphibolite sills (Fig. 4). On the southwest side of Bear Hill, one of the sills is cut by the Indian Pond pluton. They are therefore believed to represent forerunners of the pluton.

Bethlehem gneiss of the Indian Pond pluton cuts the Littleton formation and therefore must be younger than lower Devonian. The age of uraninite in pegmatite dikes associated with the Bethlehem gneiss in the Cardigan quadrangle, a few miles to the southeast, is 304 million years and thus late Devonian according to Shaub (1938, p. 339). The writer agrees with Chapman (1939, p. 148) that the intrusion of the gneiss occurred during the late Devonian.

*Haverhill granodiorite.*—Two small, stocklike masses of biotite granodiorite in the western part of Orford township are structurally and petrographically similar to a larger body mapped by Billings (unpublished map) 1½ miles south of the town of Haverhill in the Woodsville quadrangle. The granodiorite in the Mt. Cube quadrangle is light gray, medium-grained, and carries both biotite and muscovite. Associated with the northern body are aplite dikes, 50 to 100 feet wide, slightly finer grained than the main mass, and lacking biotite. A very weak foliation, more or less parallel to the regional schistosity, exists in parts of the body.

Microscopical study of the granodiorite shows a hypidiomorphic granular texture (Table 12, columns 11, 12). Plagioclase is calcic oligoclase, with abundant twinning on the albite, pericline, and Carlsbad laws; quartz is more or less strained, and locally somewhat granulated; microcline tends to be interstitial to both plagioclase and quartz. Myrmekite occurs commonly along the borders of microcline grains. Apatite and magnetite are in isolated grains, and sphene, zircon, and allanite are associated with orange-brown biotite.

Because the granodiorite clearly cuts the folds and schistosity of the enclosing rocks, and is considerably less foliated than the Bethlehem gneiss, it was probably intruded during the waning stages of the main orogeny. It therefore represents the latest intrusions of the New Hampshire magma series in the area.

*Pegmatite dikes and sills.*—Pegmatite dikes and sills, a few inches to 20 feet wide, occur in a zone a few miles wide, west of the Mt. Clough pluton, in the southeast part of the quadrangle. They are composed dominantly of quartz and potash feldspar, with very little muscovite, and, locally a small amount of biotite, garnet, or magnetite. Many of the pegmatites were intruded along the foliation planes of the Clough and Ammonoosuc formations, and others cut granodiorite of the Smarts Mountain and Mascoma groups, and Bethlehem gneiss in the Mt. Clough pluton.

The pegmatites were intruded some time after the Mt. Clough pluton, in which they may have originated. The absence of strong granulation suggests that they were not greatly affected by the principal orogeny.

### METABASALT DIKES AND SILLS OLDER THAN THE METAMORPHISM

*General statement.*—Sills and dikes of amphibolite (metabasalt) are abundant in the lower part of the Littleton formation, in the Albee formation just above the Piermont member, and locally in the Orfordville formation. In addition, an unusually large body, 400 or more feet

TABLE 14.—*Calculated chemical analysis of amphibolite sills*

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 44.18 | 46.9 | 48.80 |
| $TiO_2$ | 1.23 | 0.9 | 2.19 |
| $Al_2O_3$ | 10.34 | 14.5 | 13.98 |
| $Fe_2O_3$ | 3.02 | 3.3 | 3.59 |
| FeO | 13.90 | 10.7 | 9.78 |
| MnO | tr | | .17 |
| MgO | 11.39 | 8.3 | 6.70 |
| CaO | 12.31 | 11.3 | 9.38 |
| $Na_2O$ | 1.11 | 2.4 | 2.59 |
| $K_2O$ | 0.81 | .6 | .69 |
| $H_2O$ | 1.57 | 1.1 | 1.80 |
| $P_2O_5$ | 0.00 | | .33 |
| Total | 99.86 | 100.0 | 100.00 |

1. Hornblende from amphibolite sill in Moosilauke quadrangle (Billings, 1937, table 19).
2. Average composition of amphibolite sills, Mt. Cube quadrangle (assumed volume percent: hornblende 70, andesine ($Ab_{55}An_{45}$) 29, magnetite 1.)
3. Daly's average plateau basalt (1933, p. 17, column 60).

wide, of heterogeneous amphibolite is exposed in the bed of Jacobs Brook, 1¼ miles southeast of Orford. The sills range in thickness from 3 to 60 feet, and can be traced commonly for several hundred feet along the strike, though both ends are generally concealed. They may occupy as much as 20 per cent of a given area, or even more, as on Peaked Mountain. Where they are numerous, the sills are shown on the geological map (Pl. 4) by a special green overprint. They generally parallel the schistosity and bedding of the sedimentary rocks, though rarely they cut the structures at a slight angle. The sills are commonly folded with the sediments and have been pulled apart locally by tectonic stretching.

*Lithology and petrography.*—Most of the sills are composed of dark green, fine- or medium-grained needle-amphibolite. It is more or less schistose, although massive varieties are common in the larger bodies. The larger sills, especially in the Littleton formation, contain abundant phenocrysts of white feldspar, which have been granulated and flattened, rarely retaining a rectangular outline. They are commonly scarce or absent at the margins of the sills.

TABLE 15.—*Estimated modes of metabasalt and femic dikes and sills*

| | Meta basalt (Premetamorphism) | | | Femic rocks (Postmetamorphism) | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
| *Phenocrysts* | | | | | | | | | |
| Plagioclase | | | | 3.5 0–10 | 6.0 | | 8.0 | 1.0 | |
| Augite | | | | 0.5 0– 4 | 3.5 | | | 4.0 | |
| Chlorite and serpentine* | | | | 4.0 0–12 | 5.5 | | | 5.0 | |
| *Groundmass* | | | | | | | | | |
| Quartz | | 2.0 | 2.0 | | | | | | |
| Potash feldspar | | | | | | | | | 5.0 |
| Plagioclase | 25.5 | 27.0 | 24.0 | 54.0 45–62 | 17.0 | 54.0 | 43.0 | 26.0 | 65.0 |
| Augite | | | | 9.0 0–20 | 31.0 | 8.0 | 6.0 | 23.0 | tr |
| Brown hornblende | | | | tr 0–0.5 | tr | | | 5.0 | |
| Blue-green hornblende | 70.0 | 70.0 | 71.0 | | | | | | |
| Biotite | 3.0 | | 1.0 | 0.5 0– 4 | | 2.3 | 0.3 | | 3.0 |
| Magnetite and ilmenite | 1.5 | 1.0 | tr | 4.0 2– 6 | 1.5 | 3.5 | 3.5 | 4.0 | 3.0 |
| Apatite | | tr | | tr 0–0.5 | tr | tr | tr | | |
| Sphene | | | 1.0 | 0.5 0–1.5 | 0.5 | 2.2 | tr | 0.3 | 2.0 |
| Pyrite | | | | tr 0–tr | 1.0 | | | | |
| Chlorite and serpentine† | | | | 20.0 10–30 | 20.0 | 30.0 | 31.0 | 25.0 | 21.0 |
| Sericite | | | | a a | 7.0 | a | | 1.0 | |
| Epidote and zoisite | | | 1.0 | 2.0 0– 6 | | | | | |
| Carbonate | | | | 2.0 0–10 | 7.0 | tr | 8.0 | 6.0 | 1.0 |
| Totals | 100.0 | 100.0 | 100.0 | 100.0 | 100.0 | 100.0 | 99.8 | 100.3 | 100.0 |
| Character of plagioclase | andesine | andesine | labradorite | albite and andesine | albite | labradorite | labradorite | labradorite | andesine |
| Texture | GP | G | GP | HG P | HG P | HG | O P | PanG P | HG |
| Grain size, mm. phenocrysts | | | | 1–15 | 0.2–1 | | | | |
| Grain size, mm. groundmass | 1.0 | 1.0 | 1.5 | 0.1–1 | 0.01–0.06 | 0.05–1.0 | 1.0–1.5 | 0.25–4.0 | 0.1–0.4 |

1. Amphibolite sill in Littleton formation, 1 section.
2. Amphibolite sills in Albee formation, average of 2 sections.
3. Amphibolite dike in Orfordville formation, Jacobs Brook near Orford, 1 section.
4. Porphyritic, fine-grained gabbro, average and range of 8 sections.
5. Porphyritic, fine-grained gabbro, chilled phase 1 inch from contact, average of 2 sections.
6. Nonporphyritic, fine-grained gabbro, average of 2 sections.
7. Diabase, average of 2 sections.
8. Camptonite, average of 2 sections.
9. Syenodiorite, 1 section.

* Pseudomorphous after some ferromagnesian mineral, probably olivine.
† Probably represents altered olivine in part, augite in part.
a Abundant sericite largely in plagioclase, not estimated separately.
G = granoblastic
HG = hypidiomorphic granular
O = ophitic
P = porphyritic
PanG = panidiomorphic granular

Thin sections show that the amphibolites are composed essentially of rather uniform blue-green hornblende and plagioclase, generally in the proportion of about 7 to 3 (Table 15, columns 1 to 3). Plagioclase is generally calcic andesine, with somewhat more sodic zones at the margins of the anhedral grains. It forms a granular mosaic, commonly with poikiloblastic crystals of hornblende. Optical properties of the hornblende are given in Table 17. They are essentially similar to those of an analysed hornblende from a similar intrusive amphibolite in the Littleton-Moosilauke area (Billings, 1937, p. 556, column 20) and to the hornblende from metamorphosed volcanics in the Orfordville and Ammonoosuc formations. The analysed hornblende is an aluminous variety with much lime, and the chemical composition is shown in Table 14. In the same table are given the composition of the amphibolites of the Mt. Cube area, as calculated from the mineral composition, and of Daly's average plateau basalt (1933, p. 17, column 60). The comparison suggests that the amphibolites represent rather typical basalt, which has not undergone much change in chemical composition during metamorphism.

*Age.*—As no evidence indicates more than one period of intrusion of amphibolite sills, it is assumed that they are all younger than the early Devonian Littleton formation. Moreover, the sills are cut by the Bethlehem gneiss of the Indian Pond pluton, and by the older, fine-grained gneiss associated with it (Fig. 4). No amphibolites were observed cutting the rocks of the Oliverian magma series in the Mt. Cube area, although Billings reports such an occurrence in the Moosilauke quadrangle (1937, p. 513). The sills have certainly been involved in most of the folding and metamorphism. Evidence in the Mt. Cube area, therefore, indicates that they came into the crust just before the late Devonian orogeny, and preceded the New Hampshire magma series.

### FEMIC DIKES YOUNGER THAN THE METAMORPHISM

More than 40 femic dikes younger than the metamorphism are rather uniformly scattered throughout the area (Fig. 5). Typically they have steep dips, are 1 to 22 feet wide, and have sharp boundaries and many apophyses. They cut all of the more important rock units in the area, have suffered some deuteric alteration, but have not been affected by the folding or metamorphism. The rock types include: (1) porphyritic, fine-grained gabbro; (2) nonporphyritic, fine-grained gabbro; (3) diabase; (4) camptonite; (5) syenodiorite. The last two types occur in only 3 out of 17 dikes studied microscopically. Modes based on thin sections are given in Table 15. Probably most of them belong to the White Mountain magma series of Mississippian (?) age, although some may belong to the Triassic igneous activity.

8. Camptonite, average of 2 sections.

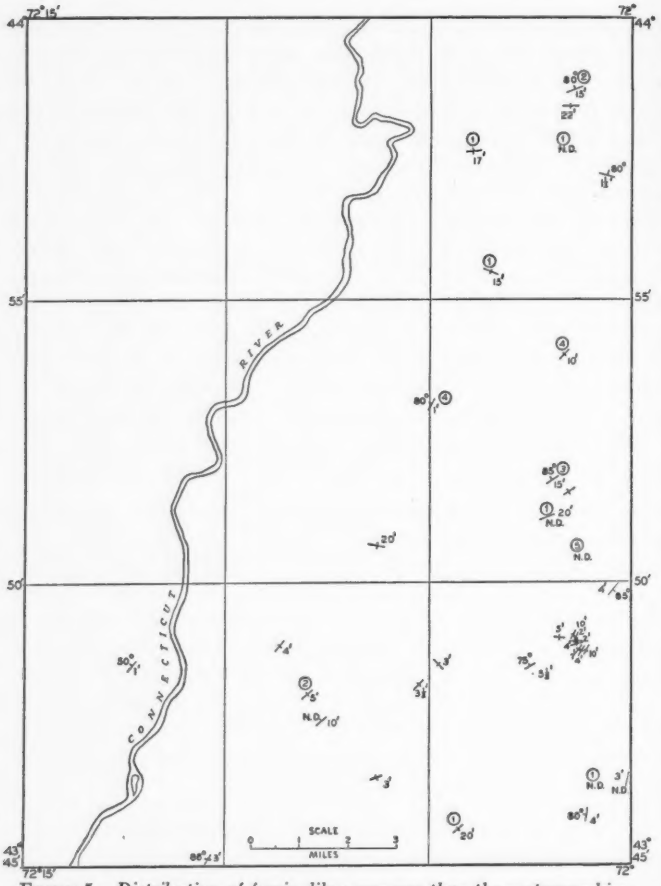


FIGURE 5.—*Distribution of femic dikes younger than the metamorphism*

Dip-strike symbol gives attitude of dike; thickness is given in feet. Numbers in circles give type of rock: 1 = porphyritic, fine-grained gabbro; 2 = nonporphyritic, fine-grained gabbro; 3 = diabase; 4 = camptonite; 5 = syenodiorite; N.D. means dip not determined.

## STRUCTURE

### GENERAL STATEMENT

The Mt. Cube area is divided by two major thrust faults into three structural belts, trending northeast-southwest (Fig. 6, Pl. 4). From northwest to southeast, the following structural units may be recognized: Piermont anticline, Ammonoosuc thrust, Salmon Hole Brook syncline, Northey Hill thrust, Bronson Hill anticline.

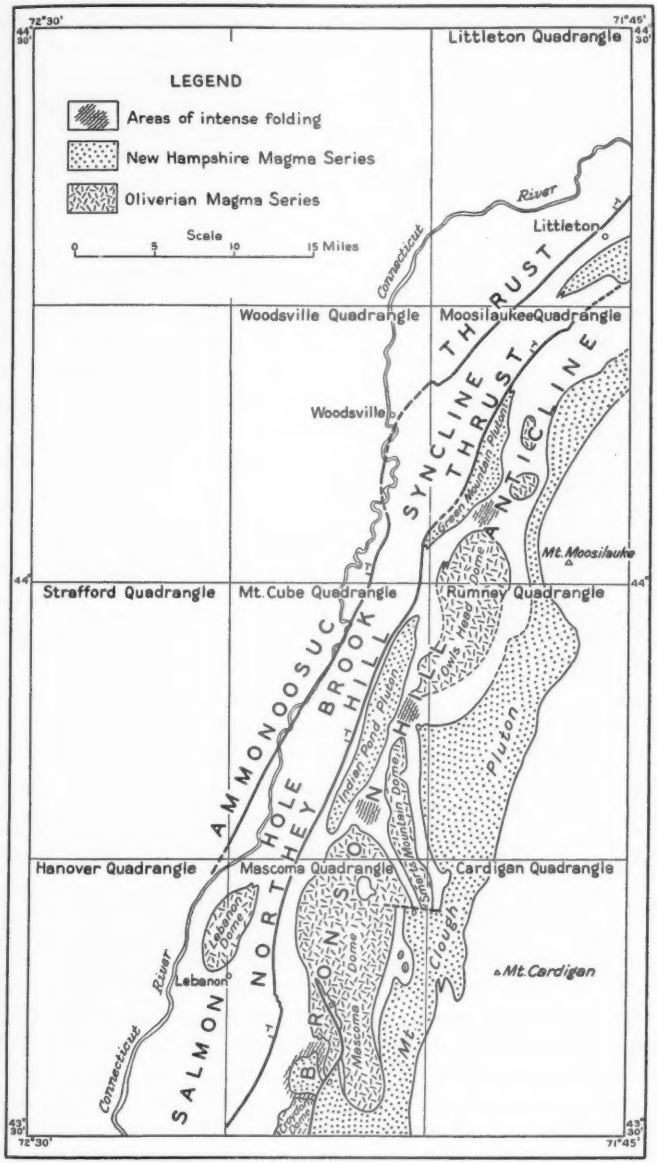


FIGURE 6.—*Major structural units of western New Hampshire*

These structural features, except the Piermont anticline continue northeastward into the Littleton-Moosilauke area (Fig. 6), where they were named by Billings (1937, p. 519 *et seq.*). Large intrusive bodies have exerted such an important influence in determining the structure of the stratified rocks in the Bronson Hill anticline, that it seems best to describe the folds, faults, and major intrusives by tectonic belts, rather than by structural types. Structural data are recorded on Figure 7.

### PIERMONT ANTICLINE

Northwest of the Ammonoosuc thrust, the Albee and Orfordville formations appear in a minor anticline, which plunges northeast, and has a large body of Fairlee quartz monzonite intruded into its southeast limb. The axial planes of minor folds are approximately parallel to the schistosity, which dips 45° to 80° SE.; the anticline as a whole is therefore overturned toward the northwest. Typical volcanics of the Sunday Mountain member were not deposited on the southeast limb of the anticline, and on the northwest limb, the Hardy Hill quartzite, here very thin, is believed to have been eliminated tectonically.

The present shape of the Fairlee quartz monzonite body is not that of the original intrusive, for it has been somewhat deformed by internal shearing and is cut off on the southeast by the Ammonoosuc thrust. The western contact of the body is well exposed and dips steeply southeast where it crosses Sawyer Mountain. At the south end, on the Palisades, the contact is essentially vertical.

### AMMONOOSUC THRUST

This fault, a major structural break in western New Hampshire, was first observed by Ross (1923) and later more fully established by Billings (1937, p. 525-530). It has now been traced from north of Littleton, New Hampshire, into the Hanover quadrangle, a distance of more than 50 miles. There are several lines of evidence for the fault in the Mt. Cube quadrangle:

(1) A distinct structural and stratigraphic break is found at the fault line. In the northern part of the quadrangle, the lower Albee formation and the Fairlee quartz monzonite, with southeastward dipping schistosity, are in contact with the Piermont member of the Albee formation, in which the bedding and schistosity dip uniformly northwest. Farther south, chlorite schists and greenstones of the Post Pond member (not differentiated on the map) of the Orfordville formation northwest of the fault, are brought against quartz-mica schist and quartzite of the same formation, southeast of the fault.

(2) There is an abrupt discontinuity of metamorphic zones at the

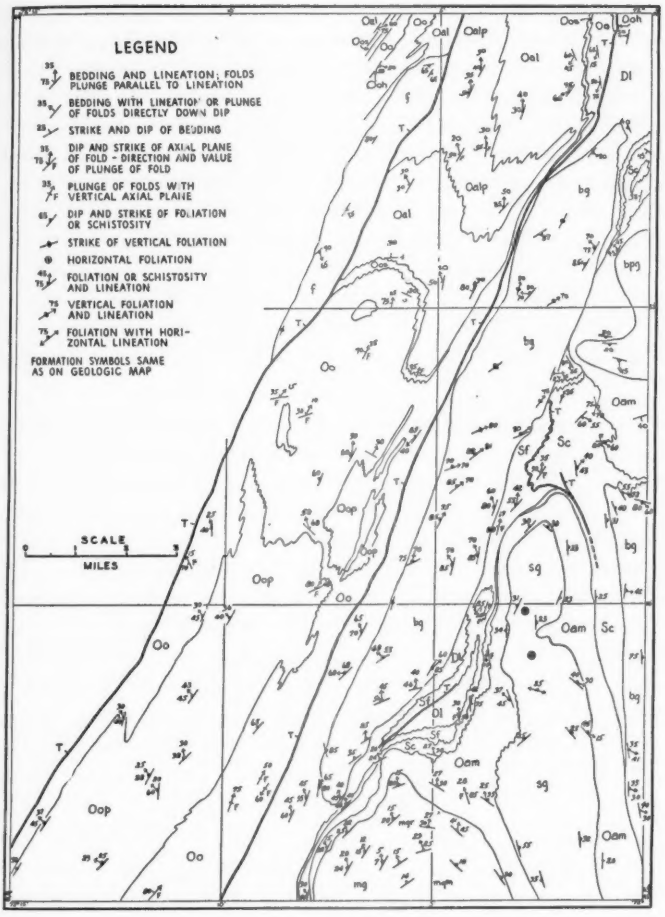
LEGEND
BEDDING AND LINEATION; FOLDS PLUNGE PARALLEL TO LINEATION
BEDDING WITH LINEATION OR PLUNGE OF FOLDS DIRECTLY DOWN DIP
STRIKE AND DIP OF BEDDING
DIP AND STRIKE OF AXIAL PLANE OF FOLD - DIRECTION AND VALUE OF PLUNGE OF FOLD
PLUNGE OF FOLDS WITH VERTICAL AXIAL PLANE
DIP AND STRIKE OF FOLIATION OR SCHISTOSITY
STRIKE OF VERTICAL FOLIATION
HORIZONTAL FOLIATION
FOLIATION OR SCHISTOSITY AND LINEATION
VERTICAL FOLIATION AND LINEATION
FOLIATION WITH HORIZONTAL LINEATION
FORMATION SYMBOLS SAME AS ON GEOLOGIC MAP
SCALE
MILES

FIGURE 7.—*Structural data for Mt. Cube area*
Formation symbols same as for geological map (Pl. 4).

fault. The rocks immediately west of the fault show only low-grade metamorphism, whereas, immediately east of it, abundant biotite, garnet, staurolite, and kyanite indicate conditions of middle-grade metamorphism. Part of the metamorphic sequence has been eliminated by the thrust. This relation is somewhat obscured in the southern part of the area by retrograde metamorphism.

(3) Although the fault plane is rarely exposed, features indicating unusual tectonic disturbance are observed in its vicinity. The systematic schistosity of the adjacent rocks becomes erratically twisted as the fault plane is approached. Brecciated and silicified rock marks the fault zone at several places. Just northeast of a tiny pond, 0.6 mile due west of the summit of Wilmot Mountain in southeastern Thetford township, a low ridge of silicified gouge, 5 to 15 feet wide and 200 feet long, occupies the fault plane. The west face of the ridge is straight and dips about 40° W.; it is believed to be approximately parallel to the fault plane. This is the only suggestion of the dip of the fault in the Mt. Cube quadrangle, but it agrees well with the dip given by Billings for the same fault further north. Seams of flinty crush rock, not hitherto described from the Ammonoosuc thrust, were seen at a few places in the underlying rocks within 100 feet of the fault.

The evidence that this is a thrust fault lies chiefly in the low dip of the fault plane, gently dipping shear planes in the overlying rocks, and in the telescoping of the metamorphic zones. Billings (1937, p. 529) says that in the Littleton-Moosilauke area the Ammonoosuc thrust has a stratigraphic separation of 5000 or 6000 feet and may have a net slip of as much as 3 miles. In the Mt. Cube area, the stratigraphic separation is probably not over 1000 feet, but the sharp contrast in the attitude of the axial planes of the minor folds on opposite sides of the fault suggests considerable horizontal movement at right angles to the strike of the fault. The writer agrees with Billings that the thrust developed late in the tectonic history of the area, after the major folding and the regional metamorphism. The fault thus cut across the folds and brought together two different metamorphic zones.

### SALMON HOLE BROOK SYNCLINE

*General features.*—This structural belt, 3 to 4½ miles wide in the Mt. Cube quadrangle, is bounded on the northwest by the Ammonoosuc thrust, and on the southeast by the Northey Hill thrust. It is dominantly a crumpled syncline, plunging northward, complicated by two subordinate anticlines.

The position of the axis of the syncline is known only from regional mapping, as the beds and the schistosity in both limbs nearly everywhere dip to the west. The axis can be traced from Peaked Mountain southwestward across Sunday Mountain, thence just west of Mud Turtle Pond, southwest of Orfordville, to a point about a quarter of a mile northwest of Whipple School. Southward from here its position is less definitely known, but it probably continues not more than half a mile west of the

Northey Hill thrust to the south border of the quadrangle. Nowhere is the synclinal axis more than a mile from the Northey Hill thrust. A large part of the eastern limb of the syncline, including the rocks below the Hardy Hill quartzite member of the Orfordville formation, has been cut out along the thrust. The excessive thinness of the upper Orfordville schist and the lowermost Albee formation northwest of Indian Pond Mountain is due, however, to the intrusion of the Indian Pond pluton.

The metamorphosed sediments in the Salmon Hole Brook syncline are everywhere much folded, and many minor folds can be seen. They are commonly isoclinal, though more open folds are not uncommon. They rarely show a systematic pattern over a considerable area, nor is their plunge consistent even in one outcrop; for both northward and southward plunges are found, ranging from nearly horizontal to nearly vertical. In the syncline as a whole, however, from Piermont village and Peaked Mountain southward to Acorn Hill in Lyme township, a predominance of folds plunging 20 to 40 degrees northward reflects the northward plunge of the main syncline. South of Acorn Hill the predominance of northward plunges is lacking, and the synclinal axis may be essentially horizontal.

The axial planes of the minor folds near the core of the syncline strike with the trend of the syncline, and are vertical or dip steeply toward the northwest. On the western limb, the northwestward dip of the axial planes becomes less, reaching 50 degrees and locally 25 degrees along the west side of the belt.

*Subsidiary folds.*—On the western ridge of Cottonstone Mountain, in Orford township, the basal Albee formation and the Sunday Mountain volcanics crop out in an anticline which plunges about 30 degrees north. Minor folds in this arcuate belt appear to be unusually open, with bedding and schistosity dipping 25 to 40 degrees northeast, north, or northwest. The anticline persists northward at least as far as the south boundary of the Piermont member of the Albee formation, but southward it disappears, and is replaced by a pronounced syncline on the south slopes of Kenyon Hill, in Lyme. Another, smaller anticline is indicated by predominantly south-plunging folds in the Albee and Orfordville formations adjacent to the Northey Hill thrust in the northeast corner of the quadrangle.

*Schistosity and cleavage.*—Axial-plane cleavage is dominant in the stratified rocks in the Salmon Hole Brook syncline, the Piermont anticline, and the adjacent part of the Bronson Hill anticline. The cleavage may be very straight, but more commonly it is wavy and locally considerably folded. Ordinarily it nearly parallels the bedding in isoclinal

folds, but in many places it cuts the bedding at large angles. The cleavage strikes about N. 30° E. and dips 40° to 85° NW. over most of the area.

An early generation of fracture cleavage, preserved by crystallization of mica and garnet porphyroblasts along the cleavage planes, is common in the Albee formation in the Salmon Hole Brook syncline. It accompanied the original folding of the more competent beds, but has been subjected to later folding and shearing so that its origin is generally obscure. It may easily be mistaken for folded and sheared bedding; and, where least disturbed, it closely resembles cross-bedding (Pl. 3, figs. 3, 4).

A later and different type of fracture cleavage is more widespread. The schistosity is folded in small chevron folds, characterized by closely spaced, sub-parallel axial planes. These axial planes, a fraction of an inch to a foot apart, represent planes of shear along which the schistosity has been bent at a sharp angle. Fracturing accompanying the bending has produced a fracture cleavage parallel to the axial planes of the folds. Commonly laminae of the schist have been dragged into the shear planes, thus producing a slip cleavage. The same sort of folding occurs very commonly without any cleavage resulting from it, and appears as minute crenulations on schistosity surfaces. This type of folding and fracture cleavage must have formed after the major recrystallization had ceased. It is distributed rather unevenly throughout the area and seems to be found where unusual shearing stresses have affected the more schistose rocks late in their tectonic history. It is especially prominent, for instance, in the vicinity of the Ammonoosuc thrust.

*Stocks of Haverhill granodiorite.*—Of the two bodies in the western part of Orford township, the more southerly is much the better exposed, and shows sharp contacts, which are commonly discordant to the local structures in the schist. The body is elongated parallel to the regional structural trend, except at its southern end, where the body consists of a series of sills connected by large crosscutting dikes. Chunky inclusions of the schist, as much as 3 feet across, are fairly common near the borders of the stock and have been considerably deformed by the intrusive. A weak foliation, manifested by biotite, is present along the eastern margin of the granodiorite; elsewhere the granodiorite is structureless.

### NORTHEY HILL THRUST

This fault, like the Ammonoosuc thrust, is an important structural break in western New Hampshire, and extends at least 20 miles north and 60 miles south of the Mt. Cube quadrangle. In the Mt. Cube area, the Northey Hill thrust has no topographic expression and is not marked

by any local distinctive features. Its existence is based upon the great stratigraphic interval between the Orfordville formation in the Salmon Hole Brook syncline to the west and the Littleton formation in the Bronson Hill anticline to the east. Stratigraphic units cut out by the fault include the upper part of the Orfordville formation, all of the Albee, Ammonoosuc, Clough, and Fitch formations, and part of the Littleton. The stratigraphic separation is therefore about 13,000 feet.

Throughout the Mt. Cube quadrangle, the rock immediately west of the thrust is black schist of the Orfordville formation, and that immediately east is very similar schist of the Littleton formation. Tracing the thrust is, therefore, difficult. At the north boundary of the quadrangle, the fault is well located because the Hardy Hill quartzite member abuts against it; and, farther south, lenses of the quartzite at intervals serve to identify the Orfordville formation.

The present attitude of the thrust, as brought out by its relation to topography, is very steep, more or less parallel to the schistosity. Because of the lack of postmetamorphic disturbance at the fault contact, it is believed that the Northey Hill thrust is older than much of the folding and metamorphism; hence the original attitude of the fault plane is in doubt. As older rocks occur west of the fault, it is considered most consistent with the general structural picture to assume that it was a shear-thrust from the west which developed early in the late Devonian orogeny.

### BRONSON HILL ANTICLINE

*General statement.*—This belt is essentially a great anticline, 6 to 10 miles wide, which is complicated by large intrusive bodies. The central part of the belt is occupied throughout the Rumney, Mt. Cube, and Mascoma quadrangles by the Owls Head, Smarts Mountain, and Mascoma domes of the Oliverian magma series (Fig. 6), surrounded by the Ammonoosuc volcanics. On the flanks of the anticline, the Clough and younger formations are exposed, and are injected by the Indian Pond and Mt. Clough plutons of the Bethlehem gneiss. Subsidiary folds and thrust faults are important locally, especially between the domes.

*Domes associated with Oliverian magma series.*—Within the Bronson Hill anticline, the metamorphosed sedimentary and volcanic rocks around the intrusives of the Oliverian magma series have taken the form of great domes or broad anticlines, called the Owls Head, Smarts Mountain, and Mascoma domes. Of these, only the Smarts Mountain and Mascoma domes are sufficiently exposed in the Mt. Cube quadrangle to warrant detailed attention here.

Nearly all of the Smarts Mountain dome lies within the quadrangle. Including its southern extension into the Mascoma quadrangle, the total

length of the dome is 12 miles, and its width 3 miles. The northern and eastern parts of this dome are relatively simple. The Ammonoosuc and Clough formations follow a remarkably smooth curve around the north end, with bedding and schistosity dipping 25 to 30 degrees away from the core. On the top of Smarts Mountain the Ammonoosuc formation appears in a cap extending nearly to the center of the dome. Along the east side the simple structure continues, with a few gentle wrinkles plunging east-northeast.

The boundary of the granodiorite core is, in general, essentially concordant with the structure of the surrounding rocks. Where the contact crosses a small valley, half a mile east-southeast of Smith Mountain, it dips 35° W. On Smarts Mountain it has an average dip of 7° E. for two thirds of a mile under the cap of Ammonoosuc volcanics (Pl. 4, section D-D′). The foliation in the margins of the intrusive closely parallels the schistosity and bedding in the surrounding rocks. Within the intrusive, the foliation is domical, as is clearly shown in a fairly well-exposed section across the dome from Smith Mountain to Smarts Mountain. This strongly suggests the former existence of a continuous cover of Ammonoosuc volcanics, now eroded everywhere except on the summit of Smarts Mountain. The position of the horizontal foliation, marking the axis of the dome, lies asymmetrically near the west side of the intrusive; and this asymmetry becomes more pronounced farther south.

Along the west flank of the dome, on Lambert Ridge, abundant minor folds appear in the Clough and Ammonoosuc formations. At the north end of the ridge, the plunge of these folds is more or less horizontal, and the axial planes are vertical or dip steeply toward the west. At the south end of the ridge and in the valley of Grant Brook, however, the folds plunge 30 to 45 degrees to the north and the axial planes dip 60 degrees or more to the east. Farther south, the west flank of the dome is poorly exposed; but on the northeast slopes of Winslow Ledge and on Sugar Hill, the Ammonoosuc volcanics dip 35° to 50° NE. In other words, the west flank of the Smarts Mountain dome is progressively overturned toward the west, beginning at Lambert Ridge, and increasing in degree toward the south.

The relations of the upper contact of the intrusive to this folded part of the cover are obscure because of poor exposures. The contact is probably folded in harmony with the cover, but it could not be definitely traced, and is highly generalized on the map. The foliation in the granodiorite, east of Lambert Ridge, is unusually weak and appears to dip gently northeast, except right at the contact. In many outcrops north of Reservoir Pond there is a linear arrangement of biotite flakes, which

is nearly horizontal and strikes N. 60° W. to N. 70° W. Farther south, east of Sugar Hill, the foliation of the core is more pronounced and near the western border it dips 50° to 60° E., in concordance with the overturned west flank of the dome.

The Mascoma dome is roughly elliptical in ground plan, measuring 6 by 17 miles, and is one of the largest of the domes so far mapped (Fig. 6). Only the north end of the dome lies in the Mt. Cube quadrangle, and this, like the north end of the Smarts Mountain dome, has a relatively simple structure. The Ammonoosuc formation wraps around the intrusive core in a smooth, semicircular arc. The schistosity and bedding in the volcanics dip away from the core at angles of 20 to 25 degrees on the north, increasing to 35 or 40 degrees on the east, and to 50 degrees on the west. The Clough, Fitch, and Littleton formations follow around the northwest flank of the dome, become somewhat more folded on the north and do not appear at all on the east flank, due to the proximity of the Smarts Mountain dome. The folds on the north slope of Winslow Ledge are open, plunging 25 to 40 degrees north or northwest (Fig. 7); the axial planes are generally vertical, but dip eastward in the bed of Grant Brook just west of Bench Mark 1110. The belt of Ammonoosuc volcanics on the east slope of Sugar Hill lies in a syncline, overturned toward the southwest, between the Smarts Mountain and Mascoma domes.

The upper contact of the intrusive core of the Mascoma dome is, in general, concordant with this domical structure in the surrounding rocks. In a valley, 600 feet deep, between Holts Ledge and Winslow Ledge, the contact may be observed, parallel with the roof rocks, dipping at an average angle of 22 degrees toward the northwest. Abundant sills of the Mascoma group occur in the Ammonoosuc formation near the contact; however, through-going dikes of the Mascoma group in the surrounding rocks were never observed.

A strong foliation is parallel to the bedding in the roof rocks of both the Mascoma and Smarts Mountain domes. The foliation is the result of the recrystallization of rocks of the Ammonoosuc, Clough, Fitch and, to some extent, the Littleton formation under stress normal to the bedding. The intensity with which these beds have been compressed and elongated to form the domes is revealed by flattened quartz pebbles in the Clough formation. In many places these pebbles are now ellipsoids, three or four times as long as thick. On the northwest flank of the Mascoma dome, where the deformation seems to have been greatest, ellipsoidal pebbles have axial ratios between 1:2:4 and 1:3:8, the shortest axis always normal to the bedding and schistosity. Assuming them to have had originally the high degree of sphericity characteristic of well-rounded

quartz pebbles (Wentworth, 1922; Graton, 1930, p. 11-12), they have been flattened normal to the bedding by at least 35 per cent, and elongated nearly 50 per cent. Thousands of pebbles, flattened and elongated all in the same direction, attest a severe, plastic thinning and stretching of the walls of the domes.

The foliation of the plutonic core of the Mascoma dome is much more pronounced than in the Smarts Mountain dome, but is similarly parallel to the general outline of the intrusive body and to the structure of the cover. Around the margins it dips outward, 20 to 30 degrees; in the interior, dips as low as 7 degrees were recorded, although nowhere in the Mt. Cube quadrangle is it actually horizontal.

The most prominent joints dip steeply toward the center of the dome. Many of them in the granodiorite have been filled with quartz-tourmaline veins. These are bordered by bleached zones an inch to several feet wide, from which the biotite of the granodiorite has been entirely removed.

*Mode of intrusion of Oliverian magma series.*—From a study of the Smarts Mountain dome and the northern part of the Mascoma dome, the writer concludes that these plutonic bodies are syntectonic phacoliths. That they are stocks or crosscutting bosses is unlikely in view of their concordant contacts and elongate shape of the Smarts Mountain granodiorite. Moreover, at least eight such bodies occur at approximately the same position in the stratigraphic sequence in western New Hampshire. They are believed rather to be sheetlike bodies, injected along a favorable horizon in the stratified rocks, and folded into anticlines or domes during the process of intrusion. The thesis that they are syntectonic is based principally upon the relation in the domes between foliated structures due primarily to the doming and linear structures which are closely related to the regional deformation.

There can be little doubt that the foliation in the roofs of the domes was caused principally by the intrusion of the cores. This must have occurred under a considerable load, probably more than the 6000 feet of known formations above the intrusives. This thickness may have been increased materially by strata deposited above the Littleton formation, and further increased in the last stages of the doming by folding and thrusting of the overlying rocks. Under this load the roof rocks were squeezed and thinned plastically as the dome developed. Recrystallization under stress essentially normal to the beds oriented mica flakes and kyanite rods, and helped to flatten and elongate quartz pebbles. The intrusion or intrusions took place so slowly that the outer parts of the intrusives crystallized and acquired a foliation also. During this time magmatic products penetrated the roof bed by bed, resulting in an

exceptionally wide zone of mixed rocks and extensive reaction between the intrusives and the country rocks.

Linear features in both roofs and intrusives were produced simultaneously with the foliation. In the roof rocks they consist of elongated quartz pebbles and mica flakes and well-oriented kyanite rods and amphibole needles. In the intrusives they consist of biotite flakes arranged in more or less prominent streaks, or elongated and oriented metacrysts. It seems clear that these linear features were formed by the recrystallization of the rocks under differential stress. Moreover, there is no evidence that this recrystallization occurred at a time different from that of the foliation, as the same minerals and structures are involved in both. Linear structures and planar structures (foliation) seem to have developed simultaneously in both roof and intrusive core.

The linear features associated with the domes are generally parallel to the axes of minor folds in the Bronson Hill anticline, and even to linear structures in adjacent parts of the Bethlehem gneiss. This is shown diagrammatically in Figure 7. As it seems unlikely that two independent sets of stresses would produce features so markedly parallel, the formation of the linear features must have accompanied the last stages of the folding. Therefore, if it is true that the foliation is a result of intrusion of the Oliverian rocks yet the foliation and lineation were produced at the same time, then the intrusion must have occurred during the folding.

Chapman (1939, pp. 155-156) regards the domes as resembling cedar-tree laccoliths, injected at considerable depth, essentially before the folding. He points to the relatively simple structure of the stratified rocks in the roofs of the domes and suggests that the roofs were largely protected from folding by the buttressing effect of the already solidified cores. On this basis, the observed linear structures would be the result of regional deformation on the domes after the intrusives had solidified.

The writer agrees concerning the depth of intrusion. He would suggest, however, that intense stress acting normal to a series of beds over a considerable area and for a long time would restrict, if not prevent, folding of those beds by regional deformation. Also, if the domes, with their concentric foliation, had been deformed by later stress sufficiently to produce the observed linear features, one would expect to find planar structures, parallel to the regional schistosity, cutting the older foliation. Where such structures should be most evident, at the north ends of the Smarts Mountain and Mascoma domes, no trace of them was found.

In conclusion, it seems probable that, although the intrusion of the Oliverian magma series may have begun early in the tectonic history of the region, it continued, at least in the sense that the intrusives remained capable of producing deformation in the surrounding rocks,

throughout the main orogeny. The intrusives, therefore, not only domed their roofs but were affected at the same time by the folding and should be classed as syntectonic phacoliths.

*Subsidiary folds and thrust faults.*—On Piermont Mountain the west flank of the Owls Head dome is complicated by an anticline in the Clough formation, with the Ammonoosuc volcanics exposed in the core. The axial planes of minor folds are vertical or dip steeply toward the west. The southwest end of this anticline has been compressed and pushed eastward by the intrusion of the Indian Pond pluton.

A subsidiary anticline on the west flank of the Smarts Mountain dome brings up a considerable mass of the Clough formation, surrounded by the Fitch, on Smith Mountain. The Fitch formation is exposed in a long, southwest continuation of the anticline, the southern part of which has been broken, and thrust southeastward over the Littleton formation.

At several places, notably on the northwest flank of the Smarts Mountain dome, and on the southwest flank of the Owls Head dome between Mt. Cube and Piermont Mountain, the Clough formation is very thin or absent, and the Ammonoosuc is in contact with the Fitch or Littleton formations. This is believed to have been accomplished by extreme tectonic squeezing and stretching, rather than by nondeposition or simple thrust faulting. A similar occurrence has been described by Billings (1937, p. 531-532).

The mass of the Clough formation on and around Mt. Cube presents a very complicated structure, because of its position between the two domes. In the southwestern part of the mass, just east of Quinttown, the Clough apparently overlies the Fitch, and both are full of minor folds plunging 30 degrees northeast. On Eastman Ledges, the Clough formation dips moderately and uniformly northeast; it represents a more or less undeformed plate which has been detached and pushed northwestward from the northeast flank of the Smarts Mountain dome. On Mt. Cube proper, most of the folds plunge steeply in various directions. The whole appears to represent an orginally saddle-shaped mass between the Smarts Mountain and Owls Head domes, which has been thrust westward over the Fitch on the north end of the Smarts Mountain dome and later crumpled. This westward movement is consonant with the overturning of the southern part of the Smarts Mountain dome, and with westward movements in the Mt. Clough pluton reported by Chapman in the Mascoma quadrangle (1939, p. 163-164) and in the western part of the Cardigan quadrangle by Fowler-Lunn and Kingsley (1937, p. 1376).

*Indian Pond and Mt. Clough plutons.*—Because little of the Mt. Clough pluton (Fig. 6) is exposed in the Mt. Cube area, and because this body

a few miles to the southeast has been described by Fowler-Lunn and Kingsley (1937) and by C. A. Chapman (1939), the present discussion will be confined to the Indian Pond pluton. The Indian Pond pluton is an essentially vertical, tabular body of granitoid gneiss, almost 10 times as long as wide, elongated parallel to the regional structure. The attitude of foliation at the border of the gneiss, and the topographic effects on the course of the contact indicate that the western contact dips from 75° NW. to vertical, except within 1 or 2 miles of the ends, where it dips 80° SE. The eastern contact is also vertical or dips very steeply northwest, except near the ends, where the dip decreases to about 45° NW. Inasmuch as the contacts at the ends of the pluton dip toward each other, and their intersection is troughlike, plunging toward the center of the body, it is believed that the length of the pluton decreases somewhat downward, as shown in structure section E-E′ of Plate 4.

The contact between the pluton and the schist is everywhere sharp, generally rather straight, and parallel to the schistosity in the surrounding rocks. Locally, blunt protuberances of the gneiss deform the schist. Where amphibolite dikes are cut by the gneiss, they may project slightly into it as though they presented greater resistance to penetration by the intrusive. At one place only does the contact cut definitely across the schistosity of the surrounding rocks; this is on cliffs on the south side of Bear Hill, at the extreme south end of the body. Here for nearly 100 feet the contact is exposed, strikes east-west, and dips about 45° N., cutting schistosity which strikes N. 60° E. and dips 45° NW. Here, too, occurs the only crosscutting apophysis of the gneiss, an aplitic dike 15 feet long and 1 foot wide. Other offshoots of the gneiss are remarkably few, and consist of sills 5 or 6 feet thick, injected along the foliation planes in the schist near the ends of the pluton.

The pluton is foliated throughout, but not uniformly. Foliation is best developed in the eastern half where it is parallel to the long axis of the pluton; along the western side it is commonly poor, except within a few feet of the contact. In the central and western parts of the pluton, where the foliation trends obliquely across the body, it is generally weak, and grades into a linear structure which plunges north. The linear structure consists largely of elongated aggregates of biotite flakes, in which the individual flakes are commonly oriented parallel to a line (lineation) rather than a plane (foliation) in the rock. Elongated quartz aggregates are also conspicuous locally in the lineation.

At the margins of the pluton, the foliation is everywhere parallel to the contacts. Along the discordant contact at the extreme south end of the pluton, however, the marginal foliation is weak, but apparently remains parallel to the axis of the pluton. Considerable areas at both

ends of the body show almost no foliation. Inclusions are not abundant anywhere in the pluton; where present they are composed of dark schist, are generally less than 1 foot long, though a few larger ones were seen, and are elongated parallel to the foliation in the gneiss.

The magma which formed the Indian Pond pluton was injected almost vertically into the schist, in which it apparently made room by forcibly pushing the country rock ahead and aside. That stoping was not important at the level now exposed is indicated by the scarcity of inclusions near the contacts, and by the paucity of dikes or sills in the country rock. Strong horizontal stress produced by the intrusion is believed to have greatly squeezed the eastern limb of the Salmon Hole Brook syncline at the north end of the pluton, and to have locally overturned the normally westward dipping schistosity. The pluton has also compressed the south end of the minor anticline which exposes the Ammonoosuc formation on Piermont Mountain.

The foliation and the commonly associated lineation throughout the pluton cannot be explained as having been produced simply by flow of a viscous magma past its walls. In fact, in many places the gneiss adjacent to the schist has relatively little of either structure. It is believed, therefore, that the foliation and lineation were caused by stresses acting from without, before complete solidification of the gneiss. The intrusion must have occurred after the development of the schistosity in the Littleton formation, as the schistosity is cut by the gneiss on Bear Hill. The remarkable parallelism of linear structures in the gneiss to linear structures and fold axes in the surrounding rocks, shows, however, that intrusion occurred before the final stages of the regional deformation.

## METAMORPHISM

### GENERAL STATEMENT

In the Littleton-Moosilauke area, Billings (1937, p. 539-544) found that the sedimentary rocks had been subjected to progressive, dynamothermal metamorphism, increasing in intensity from northwest to southeast. He divided these rocks into three zones, based upon the present mineral composition of aluminous sedimentary rocks of comparable chemical composition, as shown in column 1 of Table 16. The zones roughly parallel the trend of folds and thrust faults. The Ammonoosuc thrust sharply divides the low-grade rocks from those of middle-grade, and in general the high-grade rocks lie southeast of the Mt. Clough pluton of Bethlehem gneiss.

This relation between metamorphic zones and structural features continues essentially unchanged into the Mt. Cube area. Most of the rocks

mapped lie southeast of the Ammonoosuc thrust, and are composed of minerals characteristic of the middle-grade zone of metamorphism. In a small part of the mapped area, northwest of the thrust, rocks of the Orfordville and Albee formation belong to the low-grade zone. No rocks

TABLE 16.—*Mineralogical character of rocks in three iso-physical zones*

| | (1) | (2) | (3) |
|---|---|---|---|
| Hypothetical original rock | Shale | Dolomitic shale and sandstone | Flows and pyroclastics of basaltic or andesitic composition |
| Low-grade zone | Slate (chlorite-sericite-quartz-albite rocks) | Dolomitic slate | Albite-chlorite rocks with calcite and epidote |
| Middle-grade zone | Biotite-mica schist, with or without garnet staurolite | Actinolite-biotite schist; quartz actinolite-diopside granulite | Amphibolite: hornblende and andesine, with epidote |
| High-grade zone | Mica schist, garnet schist, sillimanite schist | (absent) | Amphibolite |

Condensed from Billings (1937, table 17, opposite p. 544).

of the high-grade zone are present in the Mt. Cube area, although they are found a few miles to the southeast in the Rumney and Cardigan quadrangles.

As the major features of the regional metamorphism in western New Hampshire have been fully discussed by Billings, the present paper will be confined to additional data on mineral relationships in the middle-grade zone and on retrograde metamorphism.

### MINERAL RELATIONS IN THE MIDDLE-GRADE ZONE

*Biotite and muscovite.*—These are stable minerals in potash-bearing rocks throughout the middle-grade zone. Muscovite is the more abundant, and always occurs in schistose rocks as small, well-oriented flakes parallel to the foliation. Porphyroblasts of muscovite are rare, although they do occur in calcareous sandstones of the Fitch formation. Biotite is largely lepidoblastic, like muscovite, but cross-biotites are conspicuously developed in schists of the Orfordville and Albee formations. These porphyroblasts are generally ellipsoidal, with their longest axes parallel to some line in the foliation plane of the rock. The unusual shape and orientation must have resulted from deformation of the enclosing rock, during or

after the development of the porphyroblasts. In some thin sections, porphyroblasts have been rotated to positions across the foliation, with attendant curving of the lepidoblastic micas around them.

*Garnet.*—Pyralspite (largely almandine) occurs in schists throughout the area. Conspicuous porphyroblasts are common; and, where these are absent, microscopic study commonly reveals abundant minute crystals. Skeletal crystals are common, with abundant inclusions of quartz, but mica flakes seem to have been absorbed by the garnet. Garnet apparently formed later than most of the mica, but sheared crystals show that it did not wholly escape the deformation.

*Staurolite.*—In the western part of the belt between the Ammonoosuc and Northey Hill thrusts, staurolite is characteristic of the schists of the Orfordville formation and the Piermont member of the Albee formation. It also appears in the Littleton formation on Mt. Cube, and on the west slope of Lambert Ridge, as well as in mica schist of the Clough formation in the vicinity of Mt. Cube. In the Orfordville and Piermont rocks, the staurolite is concentrated in certain beds, and absent in intervening ones, indicating considerable variation in the ratio of alumina to other oxides in different strata. The absence of staurolite from a large part of the schist in the Orfordville and Littleton formations is probably due to insufficient alumina, except for a small area of chloritoid schist. Staurolite porphyroblasts are not fractured, nor are they systematically oriented; thus they escaped the major deformation. In many places, especially near the Ammonoosuc thrust, they are wholly or partially replaced by a mixture of muscovite and chlorite, resulting from retrograde metamorphism.

*Kyanite.*—This mineral is associated with staurolite in many places in the Mt. Cube area, and is not commonly found otherwise. It appears to be entirely stable, and to represent excess alumina beyond the available FeO necessary to make staurolite. Kyanite is even more susceptible than staurolite to retrograde metamorphism and is generally altered to a pseudomorph of muscovite.

*Ottrelite.*—Rectangular porphyroblasts of ottrelite as much as 1.5 millimeters long occur in a narrow zone of carbonaceous, quartz-muscovite schist southeast of Sunday Mountain in Orford township. Biotite and garnet occur in the adjacent schists, but staurolite does not. In view of the close chemical similarity between chloritoid and staurolite, and the fact that they are not known to occur together, it may be argued that they do not exist together in equilibrium. Inasmuch as chloritoid is

known to occur in slates, the chloritoid-bearing rocks in the vicinity of the Northey Hill thrust may be in an area where the temperature did not become as high as in the area further west.

*Chlorite.*—This mineral, normally characteristic of the low-grade zone, is widespread in the rocks of the middle-grade zone in the Mt. Cube area. Most of it has resulted from retrograde metamorphism, but under some conditions chlorite appears to be stable in the middle-grade zone.

Chlorite due to retrograde metamorphism (penninite or ripidolite) surrounds porphyroblasts of biotite, garnet, and staurolite, fills cracks in them, or replaces them altogether. Where chlorite has replaced biotite, especially in igneous rocks, it commonly contains much rutile and leucoxene. Chlorite is especially abundant in schists of the Littleton formation, west of the Indian Pond pluton. It occurs in fresh, clear-cut flakes associated with muscovite and biotite, closely resembling and partly taking the place of the latter. Optical data indicate that the chlorite is an aluminous variety with about as much iron as magnesium, probably ripidolite or diabantite (Winchell, 1933, p. 280). Normally such chlorite in the middle-grade zone should react with muscovite to form biotite. These rocks, therefore, must represent a state of disequilibrium, in which the reactions did not go to completion. The similarity of the habit of the chlorite to that of the biotite, and the presence of partially chloritized garnets in some of these rocks, suggest that chlorite was being formed at the expense of biotite when the reactions ceased.

The only variety of chlorite which appears to be stable in the rocks of the middle-grade zone is clinochlore or pro-chlorite. It occurs in large porphyroblasts enclosing hornblende needles in amphibolite in the Ammonoosuc and Post Pond volcanics. Given an abundance of iron and magnesium, and a deficiency of potash, this chlorite will form well into the middle-grade zone.

*Plagioclase.*—The most common plagioclase observed is andesine or oligoclase, although labradorite occurs in the more calcic rocks. In rocks deficient in lime, albite is the stable plagioclase. Zoned plagioclase, with calcic cores, occurs to some extent in metamorphosed volcanic and intrusive rocks. Most of the zoned crystals were probably inherited from the original rocks. In several instances, however, reversed zoning unrelated to crystal outlines may have been produced by metamorphic reactions.

*Epidote.*—Excess lime is commonly found in epidote (clinozoisite or pistacite with about 10 per cent ferrous oxide), especially in the more femic rocks. It is prominent in calcareous hornblende gneiss of the Post

Pond volcanics, and locally in the Ammonoosuc volcanics it amounts to 80 per cent of the rock. Some epidote has resulted from metamorphism of labradorite, or more calcic plagioclase, and is scattered through plagioclase crystals. Most of it is independent of the feldspars, however.

TABLE 17.—*Optical properties of hornblendes in metamorphosed basaltic rocks*

| (1) | (2) | (3) | (4) |
|---|---|---|---|
| $\alpha = 1.651$<br>$\beta = 1.665$<br>$\gamma = 1.672$ | 1.649<br>1.662<br>1.672 | 1.656<br>1.669<br>1.675 | 1.654<br>1.667<br>1.676 |
| X = pale yellow<br>Y = yellow green<br>Z = blue green | pale yellow<br>grayish green<br>blue green | lemon yellow<br>grass green<br>greenish blue | yellow green<br>yellow green<br>blue green |
| $Z \wedge C$ 20° | 18° | 19° | 19° |

(1) From amphibolite "pillow" in Post Pond volcanics; Hewes Brook, Lyme township, Mt. Cube quadrangle.
(2) From amphibolite sill, 0.8 mi. northeast of Bench Mark 805, Piermont township, Mt. Cube quadrangle.
(3) From amphibolite sill, summit of Peaked Mountain, Piermont township, Mt. Cube quadrangle.
(4) Analysed hornblende from amphibolite sill, Sugar Hill, Moosilauke quadrangle. Data from Billings (1937, p. 513).

Clinozoisite is abundant, along with other lime-silicate minerals, in metamorphosed dolomitic rocks of the Fitch formation.

*Hornblende.*—Blue-green, aluminous hornblende is as widespread in the femic rocks of the Mt. Cube area as is muscovite in the potash-rich rocks. It is the principal constituent of metamorphosed flows, pyroclastics, and intrusives of basaltic composition. Optical data on several hornblendes from the Mt. Cube quadrangle compared with those of an analysed specimen from the Moosilauke quadrangle (Table 17) show marked similarity in all of them.

*Tremolite, actinolite, diopside.*—Tremolite and actinolite occur chiefly in metamorphosed calcareous sediments of the Fitch formation. They are associated with diopside, clinozoisite, intermediate plagioclase, and potash feldspar in an unusual series of granulites in which micas are conspicuously absent. The most widespread of these is an actinolite-feldspar rock representing dolomitic shale or sandstone at the top of the Fitch formation. In other, less common types, diopside and an orange garnet, largely grossularite, take the place of actinolite and plagioclase.

In the reaction of dolomite with quartz to form tremolite, in ordinary dynamothermal metamorphism, considerable calcium carbonate is left

over, and this combines with tremolite at a somewhat higher temperature to form diopside. In these rocks, however, the carbon dioxide has escaped, and the lime has gone into plagioclase. This reaction requires considerable alumina, which must have come from original clay minerals or from

TABLE 18.—*Mineral composition of actinolite-feldspar granulite and inferred composition of the original sediments*

| | (1) | (2) | (3) | (4) |
|---|---|---|---|---|
| Quartz | 6 | 8 | 33 | 35 |
| Potash feldspar | 4 | 35 | | 10 |
| Plagioclase | 35 | 32 | | |
| Actinolite | 40 | 25 | | |
| Clinozoisite | 10 | | | |
| Dolomite | | | 31 | 21 |
| Calcite | 5 | | 4 | |
| Sericite | | | 4 | 31 |
| Chlorite | | | 10 | 3 |
| Kaolinite | | | 18 | |

(1) Actinolite-granulite from the quadrangle boundary, north slope Piermont Mountain.
(2) Actinolite-granulite from 1.8 mile N. 78° E. of Lyme Center.
(3) Hypothetical dolomitic, argillaceous sandstone; composition calculated from column 1.
(4) Hypothetical dolomitic, arenaceous shale; composition calculated from column 2.

sericite. In granulites with much potash feldspar, it seems probable that sericite has reacted with dolomite to form tremolite, anorthite and orthoclase, with the liberation of carbon dioxide. Where little potash feldspar is present, the alumina must have come from original clay minerals. Although some soda and possibly potash may have been introduced into these rocks from without, the persistence of the actinolite-feldspar granulites at the same stratigraphic position throughout the quadrangle suggests that their original composition has been the controlling factor in their present mineral content.

### RETROGRADE METAMORPHISM

Mention has been made of sporadic development of chlorite from minerals of a higher grade of metamorphism. This is the result of reactions during the time when the temperature of the rocks was falling, after the climax of dynamothermal metamorphism. Rarely are large areas affected, and in general the time during which such reactions could occur has been so short, and the reactions probably so slow, that they have not gone to completion. Thus many of the middle-grade minerals remain.

A notable exception to this is an area in the southwestern part of the quadrangle, immediately southeast of the Ammonoosuc thrust. Here quartzites and quartz-mica schist of the Orfordville formation have been extensively altered. This has been accompanied by a considerable development of feldspar and obliteration of the schistosity, so that these rocks have a very different appearance from normal schists of the Orfordville formation. They are composed chiefly of quartz, feldspar (much of it orthoclase), chlorite, and muscovite. Most of the chlorite has been derived from biotite, for pseudomorphs of biotite porphyroblasts are very common. Biotite has been entirely replaced as far as 1000 feet southeast of the thrust. Garnets have been completely altered to chlorite for 300 feet from the thrust; staurolite and kyanite, normally abundant in this part of the Orfordville formation, are represented only by pseudomorphs for distances as great as 800 feet.

The southeastern boundary of the zone of complete chloritization is very irregular. Rocks in which all the biotite has been altered occur as much as 2000 feet from the thrust in some places, and remnants of biotite persist to within 100 feet at other places. The average width of the zone of nearly complete alteration is about 1000 feet. Although rocks adjacent to the Ammonoosuc thrust farther north are not well exposed, similar effects can be seen, especially southwest of Piermont village.

The presence of potash feldspar is unusual as a product of retrograde metamorphism. Although some of it may be inherited from the original sediments, study of thin sections indicates that potash, liberated by the chloritization of biotite, has reacted with muscovite and quartz to form feldspar. Two facts which tend to substantiate this view are: (1) where there is abundant feldspar muscovite is scarce, and vice versa; (2) quartz grains are commonly embayed by feldspar grains. Both of these features suggest that feldspar has formed at the expense of quartz and muscovite.

The origin of this belt of intensely altered rocks may be sought in three possible factors: (1) spontaneous reactions at a lower temperature than that at which the middle-grade minerals were formed; (2) reactions effected by hydrothermal solutions rising along the thrust zone, and migrating eastward along westward-dipping foliation and bedding planes in the footwall of the fault; (3) reactions promoted by strain due to movement on the thrust plane. The concentration of the alteration in rocks adjacent to the thrust clearly points to the second and third factors. Within a few hundred feet of the fault, excessive strain is everywhere evident. Distorted schistosity, closely spaced shear planes, and shredded micas are characteristic features, and thin seams of ultramylonite were found in several places. That hydrothermal solutions were important is indicated by the irregular boundary of the altered zone, and by marked

alteration relatively far from the thrust plane, especially in rocks which have not been excessively sheared. No great changes in chemical composition seem to have resulted, however.

## GEOLOGICAL HISTORY AND CONCLUSIONS

The recorded geological history of the Mt. Cube area comprises three major stages: (1) geosynclinal sedimentation, accompanied by volcanism, from middle Ordovician (?) through lower Devonian; (2) a comparatively short, but violent orogenic revolution, accompanied by extensive intrusions, during the late Devonian; (3) a long period of erosion from the late Devonian to the present, although there may have been unrecorded periods of volcanism or sedimentation.

During the earliest recorded stages of the geosynclinal phase the middle Ordovician (?) Orfordville formation was deposited. The accumulation of carbonaceous marine mud and sand was accompanied by extensive volcanic activity, particularly submarine basaltic eruptions. Altogether sediments and volcanics totaling 5000 feet were deposited. Subsequently, during the upper Ordovician (?) 5000 feet of sand, mud, and a little quartzose gravel was laid down to form the Albee formation. Extensive volcanism then ensued and the 2000 feet of quartz latite, dacite, and basalt that constitute the Ammonoosuc volcanics was deposited, chiefly as pyroclastic material.

Sedimentation was interrupted in the late Ordovician (?) by mild orogeny, the intrusion of the Fairlee quartz monzonite, and subsequent erosion.

As the early Silurian seas transgressed over the older rocks, sand, argillaceous sand, and quartzose gravel, reaching a maximum probable thickness of 1200 feet, accumulated to form the Clough formation. In the middle Silurian the Fitch formation was deposited. Calcareous ooze, mud, dolomitic sand, and dolomitic mud, with a probable original thickness of 400 to 600 feet, gathered on the sea floor. Toward the south the Fitch apparently interfingers with the upper part of the Clough formation. During the lower Devonian, mud and argillaceous sand of the Littleton formation were deposited. Although only 3000 feet remains in the Mt. Cube area, farther north, in the Littleton-Moosilauke area, the maximum thickness reaches 5000 feet.

The great orogeny was late Devonian. The Ordovician (?), Silurian, and Devonian strata were folded, faulted, metamorphosed, and injected by igneous rocks. The resulting structures trend north and northeast. A major arch, the Bronson Hill anticline, rose in the eastern part of the Mt. Cube area. On the northwestern flank of the arch a master fault, the Northey Hill thrust, developed and along it the Ordovician (?) rocks

to the west were thrust upward. The rocks west of this fault are folded into a major syncline, on which countless minor folds are superposed. The axis of this syncline lies half a mile northwest of the thrust. At a late stage in the orogeny the rocks in the northwestern part of the Mt. Cube area were driven southeastward along the Ammonoosuc thrust, perhaps several miles.

Intrusive bodies, both large and small, were related to the orogeny. Basaltic dikes and sills, now amphibolite, were injected just before or during the early stages of the orogeny. The large, concordant, intrusive domes of the Oliverian magma series, exposed in the center of the Bronson Hill anticline, have been considered to be pretectonic by Billings (1937) and C. A. Chapman (1939). The writer believes, however, that these intrusives were still partly liquid during the height of the orogeny and are syntectonic. This conclusion is based on the fact that in the Oliverian magma series a lineation, which is considered primary, is parallel to the fold axes in the adjacent schists. The large Indian Pond pluton of Bethlehem gneiss was injected during the later stages of the folding, a conclusion likewise based in part on the fact that the lineation is parallel to the fold axes of the adjacent schists. Two small bodies of crosscutting Haverhill granodiorite were intruded after most of the orogenic stress had ceased.

The orogeny was accompanied by progressive dynamothermal metamorphism, the intensity of which generally increases from northwest to southeast. The rocks northwest of the Ammonoosuc thrust belong to the low-grade zone, those to the southeast belong to the middle-grade zone. The paragenesis in the middle-grade zone is similar to that in the Littleton-Moosilauke area, and such characteristic minerals as biotite, muscovite, garnet, staurolite, kyanite, plagioclase, blue-green hornblende, tremolite, actinolite, and diopside are well displayed in the Mt. Cube area. Ottrelite is found in a small area southeast of Sunday Mountain, where the temperature was not so great as elsewhere in the middle-grade zone. Chlorite, either clinochlore or prochlorite, appears to have been locally a stable mineral in potash-deficient rocks in the middle-grade zone. Potash feldspar in the Fitch formation is considered to be due to the reaction of dolomite and sericite.

Some of the porphyroblasts, such as biotite and garnet, are rotated or sheared, which indicates that they formed before the orogeny ceased. Others, such as staurolite, are unoriented and unfractured, indicating that they escaped the major deformation,

Along the Ammonoosuc thrust the zones of progressive metamorphism were displaced. Moreover, just east of this fault a belt of extensive retrograde metamorphism is 100 to 2000 feet wide. In this belt biotite,

garnet, staurolite, and kyanite have been destroyed and the rocks now consist of quartz, orthoclase, chlorite, and muscovite. Although some of the orthoclase may be a relic from the original sediments, much of it is a product of the retrograde metamorphism. Hydrothermal solutions, which moved along the thrust plane and spread eastward along westward-dipping foliation planes, caused the retrograde metamorphism.

Femic dikes, younger than the orogeny, are probably Mississippian (?) and Triassic. The Mississippian (?) Moat volcanics may have covered the Mt. Cube area and the Triassic continental deposits may have extended this far north. In general, however, the Mt. Cube area has been subjected to erosion since the close of the Devonian. The evidence for the later stages in this erosion is preserved, but this subject, as well as the effects of Pleistocene glaciation, are beyond the scope of this paper.

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Geological Museum, Oxford Street, Cambridge, Mass.

Manuscript received by the Secretary of the Society, February 4, 1941.

STATE OF NEW HAMPSHIRE, HIGHWAY DEPARTMENT
FREDERIC E. EVERETT, COMMISSIONER AND STATE ENGIN

RTMENT
E ENGINEER

NEW HAMPSH
MT. CUBE Q

NEW HAMPSHIRE-VERMONT
MT. CUBE QUADRANGLE

(Moosilauke)

72°00'
44°00'

A'

B'

C

(Rumney)

C'

55'

50

Brook Road Sch.

Naked Mtn

Hollow

Indian Pond

Indian Pd.

Mt. Cube House

Mt. Cube Sch.

Mt. Cube

Quintstown

Eastman Ledges

Black Hill

Lemster Hill

YOUNGER THAN LOWER DEVONIAN, PROBABLY LATE DEVONIAN

DEVONIAN

Oliverian Magma Series

New Hampshire Magma Series

Lower

SILURIAN

Middle

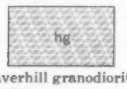
Lower or Middle

Late Ordovician?

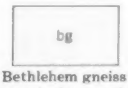
# LEGEND

## SEDIMENTARY, IGNEOUS, AND METAMORPHIC ROCKS

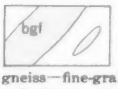
*New Hampshire Magma Series*



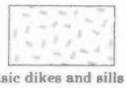

**Haverhill granodiorite**
*(Medium-grained, white or gray granodiorite, locally weakly foliated; composed of quartz, calcic oligoclase, microcline, biotite, and muscovite.)*




**Bethlehem gneiss**
*(Medium- to coarse-grained gray orthogneiss composed of quartz, oligoclase or andesine, microcline, biotite, and muscovite, commonly with phenocrysts of microcline; usually quartz-monzonite, locally granodiorite or granite.)*

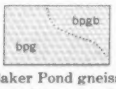



**Bethlehem gneiss—fine-grained phase**
*(Fine- or medium-grained gray gneiss, composed of quartz, oligoclase, microcline, biotite, and a little muscovite; usually quartz monzonite, locally granodiorite or granite.)*

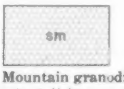


**Areas where basic dikes and sills are abundant**
*(Fine- to medium-grained, dark-green amphibolite, commonly porphyritic, composed of hornblende, andesine, and locally biotite.)*

*Oliverian Magma Series*



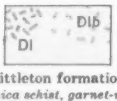

**Baker Pond gneiss**
*(Fine-grained, gray gneiss, composed of andesine, quartz, microcline, biotite, and epidote, locally with phenocrysts of microcline; largely quartz monzonite with some granodiorite. Border phase, bpgb: dark greenish gray, quartz diorite gneiss, composed of saussuritized andesine, green hornblende, biotite, potash feldspar, quartz, and epidote.)*




**Smarts Mountain granodiorite**
*(Fine- to medium-grained, light-gray, weakly foliated granodiorite; composed of oligoclase, quartz, microcline, biotite, and a little magnetite.)*

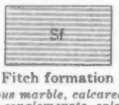
**Mascoma group**
*(Fine- to medium-grained, white, gray, or pink granitoid rocks, generally well foliated; composed of oligoclase or andesine, quartz, microcline, and biotite, locally hornblende. mpd, quartz diorite; mgr, oligoclase granodiorite; mqm, quartz monzonite; mg, granite.)*

*Lower*



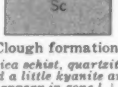

**Littleton formation**
*(Zone m: quartz-mica schist, garnet-mica schist and staurolite schist. Does not appear in zone l. Dlb, areas intruded by dikes and sills of Bethlehem gneiss, fine-grained phase.)*

*Middle*




**Fitch formation**
*(Zone m: Arenaceous marble, calcareous quartzite, quartzite, quartz conglomerate, calcareous biotite schist, mica schist, and lime-silicate granulites. Does not appear in zone l.)*

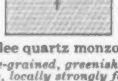
*Lower or Middle*




**Clough formation**
*(Zone m: quartz-mica schist, quartzite, and quartz conglomerate, and a little kyanite and staurolite schist. Does not appear in zone l.)*

*UNCONFORMITY*

*Late Ordovician?*




**Fairlee quartz monzonite**
*(Medium- to coarse-grained, greenish gray or pinkish quartz monzonite, locally strongly foliated; composed of bluish quartz, microperthite, saussuritized plagioclase, and green, sagenitic biotite.)*

## CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

Indefinite as sharp contact is lacking

## SPECIAL SYMBOLS

*(Dip and strike symbols representing only a small percentage of the field observations)*

75°

Strike and dip of bedding, including inverted and normal strata

Strike of vertical beds

75°

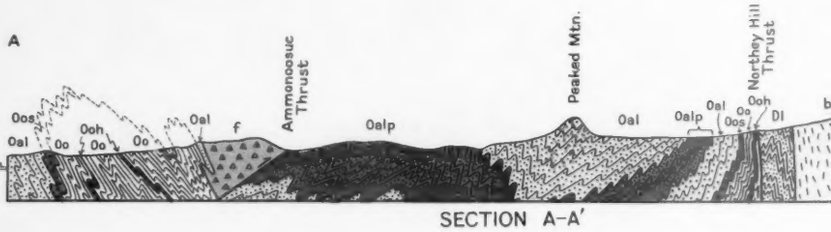
Strike and dip of foliation and schistosity

Strike of vertical foliation and schistosity

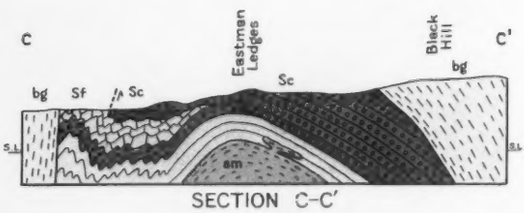
Horizontal foliation and schistosity

Overthrust side of thrust faults

Mines, prospects, and quarries mostly abandoned




SECTION A–A'

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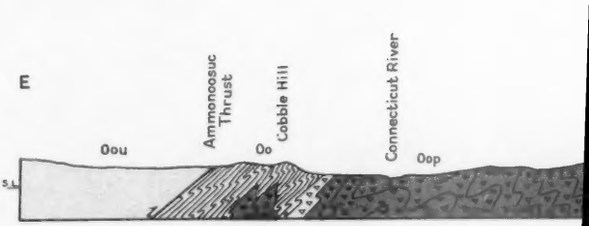
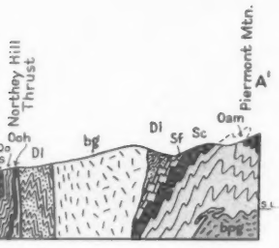
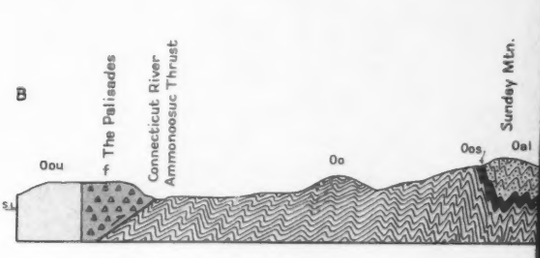
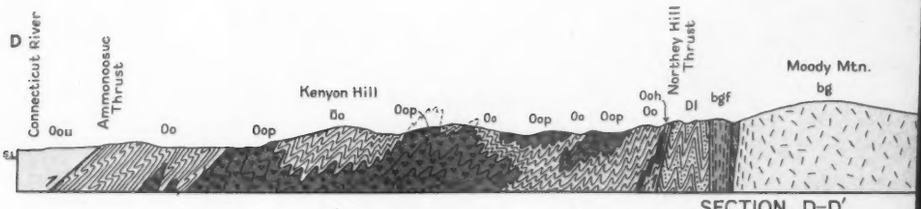



# GEOLOGIC MAP AND STRUCTURE S

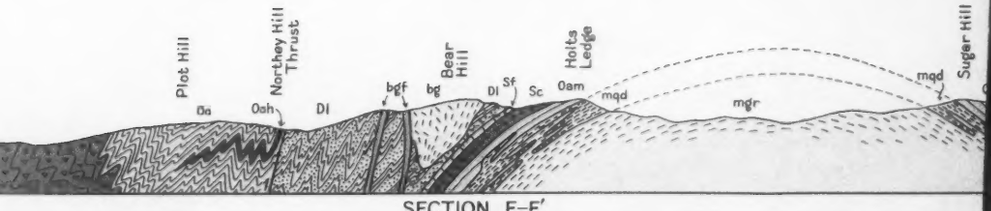
Topographic base by U. S. Geological Survey, surveyed in cooperation with the State of New Hampshire.




SECTION D–D'




SECTION E–E'

# RE SECTIONS OF THE NEW HAMPSHIRE PORTION OF

Scale $\frac{1}{62500}$

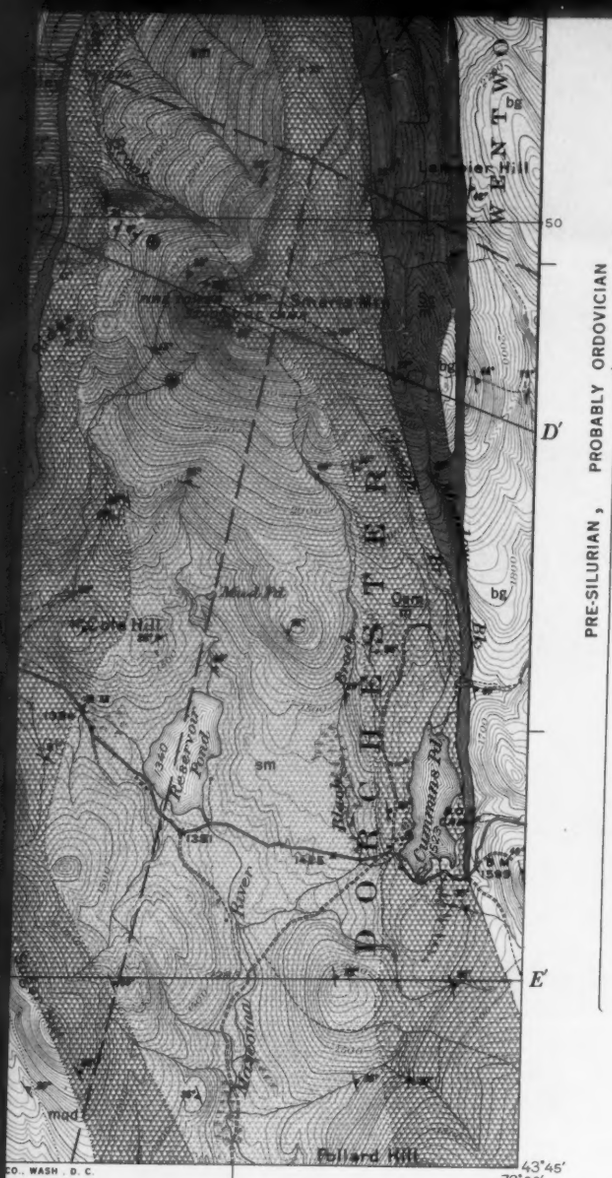
1 0 1 2 3 4 Miles

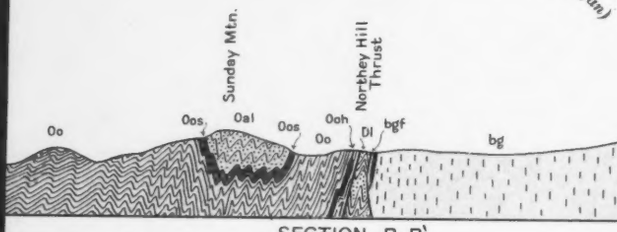
5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 5 Kilometers

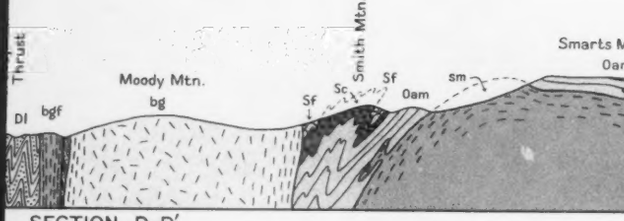
Contour interval 20 feet

*Datum is mean sea level*




SECTION B–B'

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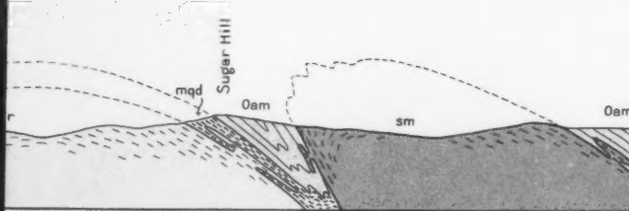
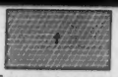



# PORTION OF THE MT. CUBE QUADRANG

4 Miles
20000 Feet
5 Kilometers

Geology by Jarvis B. Hadley, assi
John T. Hack, Walter S. White, a
Directed by Ma·land Billings. C
and 1935, under the auspices of
Sciences, Harvard University, wit
the Milton Fund. Published 19

Late Ordovician?



**Fairlee quartz monzonite**

*(Medium- to coarse-grained, greenish gray or pinkish quartz monzonite, locally strongly foliated; composed of bluish quartz, microperthite, saussuritized plagioclase, and green, sagenitic biotite.)*

Upper Ordovician?

**Ammonoosuc volcanics**

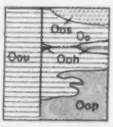
*(Zone m: fine-grained biotite gneiss, amphibolite and amphibole gneiss. Does not appear in zone l.*

Lower part of Ammonoosuc volcanics and upper part of Albee formation, which are present elsewhere, are missing in this area due to the Northey Hill thrust.

**Albee formation**

*(Zone m: thin-bedded quartzite, garnetiferous quartz-mica schist and mica schist. Fiermont member, Oalp, consists of above types with interbedded dark-gray quartz-mica schist, mica schist, and staurolite schist, and local beds of fine conglomerate in part of volcanic origin. Zone l: thin-bedded quartzite, argillaceous quartzite and light gray slate. Fiermont member, Oalp, does not appear in zone l.)*

Middle Ordovician?




**Orfordville formation**

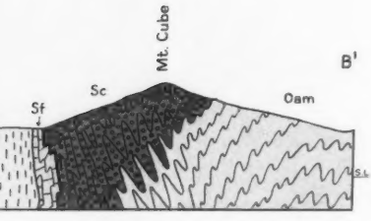
*(Zone m: black to dark-gray mica schist, quartz-mica schist, feldspathic quartzite, staurolite schist and staurolite-kyanite schist; locally otirelite schist or black, schistose, graphitic quartzite. Lenses of massive or thinly bedded, fine-grained biotite gneiss of volcanic origin in upper part. Sunday Mtn. volcanic member, Oos, consists chiefly of fine- to medium-grained amphibolite, with some fine-grained biotite gneiss and black, graphitic quartzite. Hardy Hill quartzite member, Ooh, consists of gray to white, massive quartzite and quartz conglomerate. Where the Hardy Hill is very thin or absent, its horizon is shown by a dash-dot line. Post Pond volcanic member, Oop, consists of amphibolite, amphibole gneiss, calcareous amphibole gneiss, fine-grained biotite gneiss and gray quartz-mica schist. Zone l: dark gray slate. Sunday Mtn. volcanic member, Oos, consists of schistose greenstones. Hardy Hill quartzite member, Ooh, consists of greenish gray quartz conglomerate with sericitic matrix. Post Pond volcanic member, Oop, consists of green chlorite-sericite schist. In Vermont, west of the Ammonoosuc Thrust, rocks of low-grade type, Oou, believed to be part of the Orfordville formation, have not been differentiated.)*

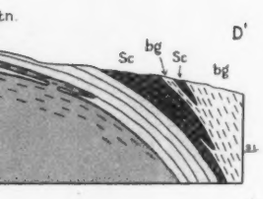
## METAMORPHIC ZONES

Shown below formation symbol thus: $\frac{Dl}{m}$

l-low grade

m-middle grade



# RANGLE

Hadley, assisted by Charles S. Denny, S. White, and Robert A. Williams. Billings. Geology surveyed in 1934 auspices of the Division of Geological versity, with the aid of grants from blished 1938.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 53, PP. 177-178, 1 PL. JANUARY 1, 1942

# GEOLOGICAL MAP OF THE CARDIGAN QUADRANGLE, NEW HAMPSHIRE

BY KATHARINE FOWLER-BILLINGS

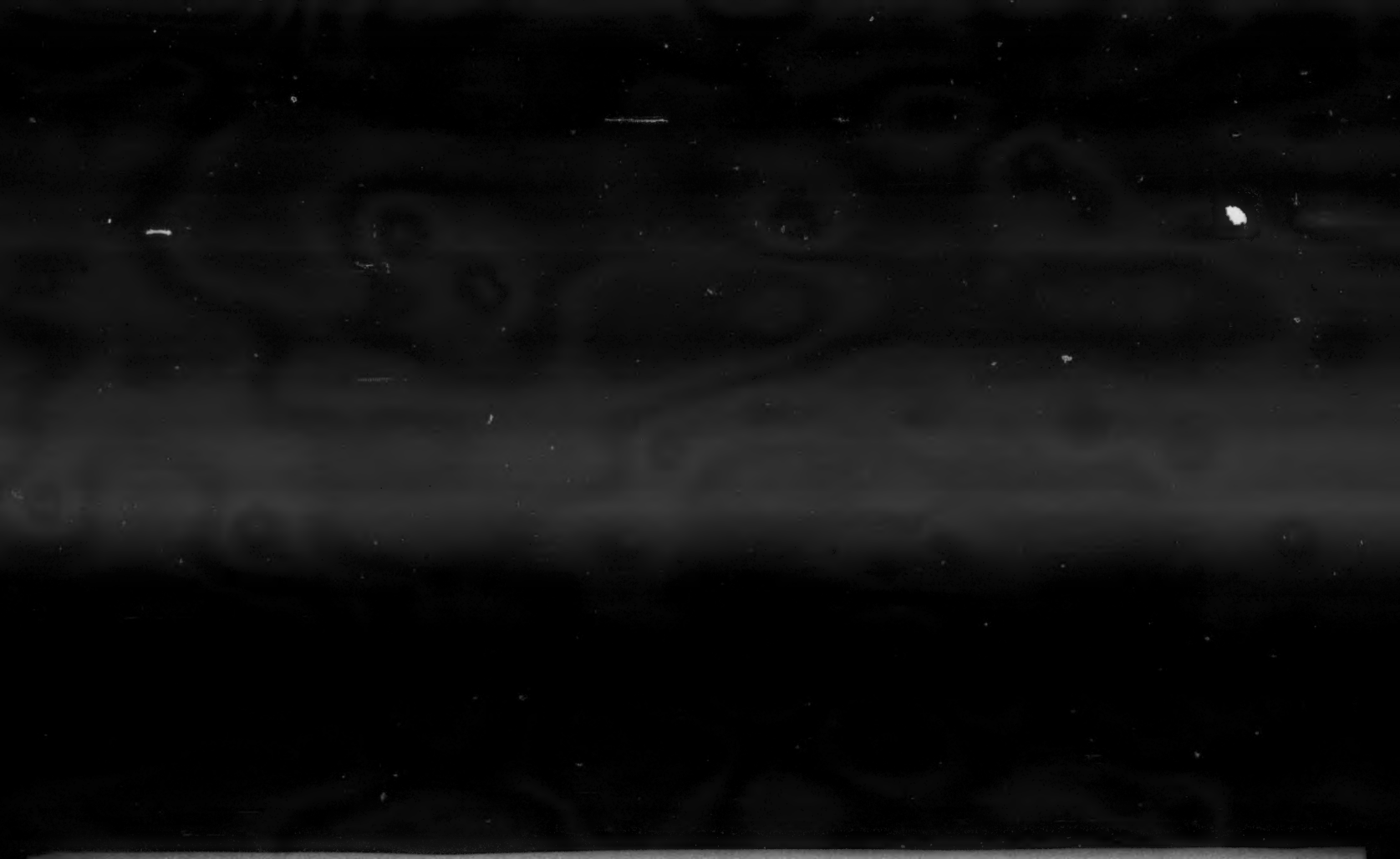
## EXPLANATION

Three years ago a paper describing the geology of the Cardigan quadrangle, New Hampshire, was accompanied by a geological map in black and white (Fowler-Lunn and Kingsley, 1937, Pl. 3). This map was on a scale of approximately 2 miles to the inch, showed no topography, and had very little of the culture. Although entirely satisfactory to illustrate the major features of the geology, the map is not serviceable in the field, nor does it bring out the relation between lithology and topography.

In the spring of 1940 the State Highway Department of New Hampshire provided funds for engraving and printing a colored geological map of the Cardigan quadrangle. The material for this map was assembled by the senior author of the original paper and is published here in the hope that it may be a service.

## REFERENCE CITED

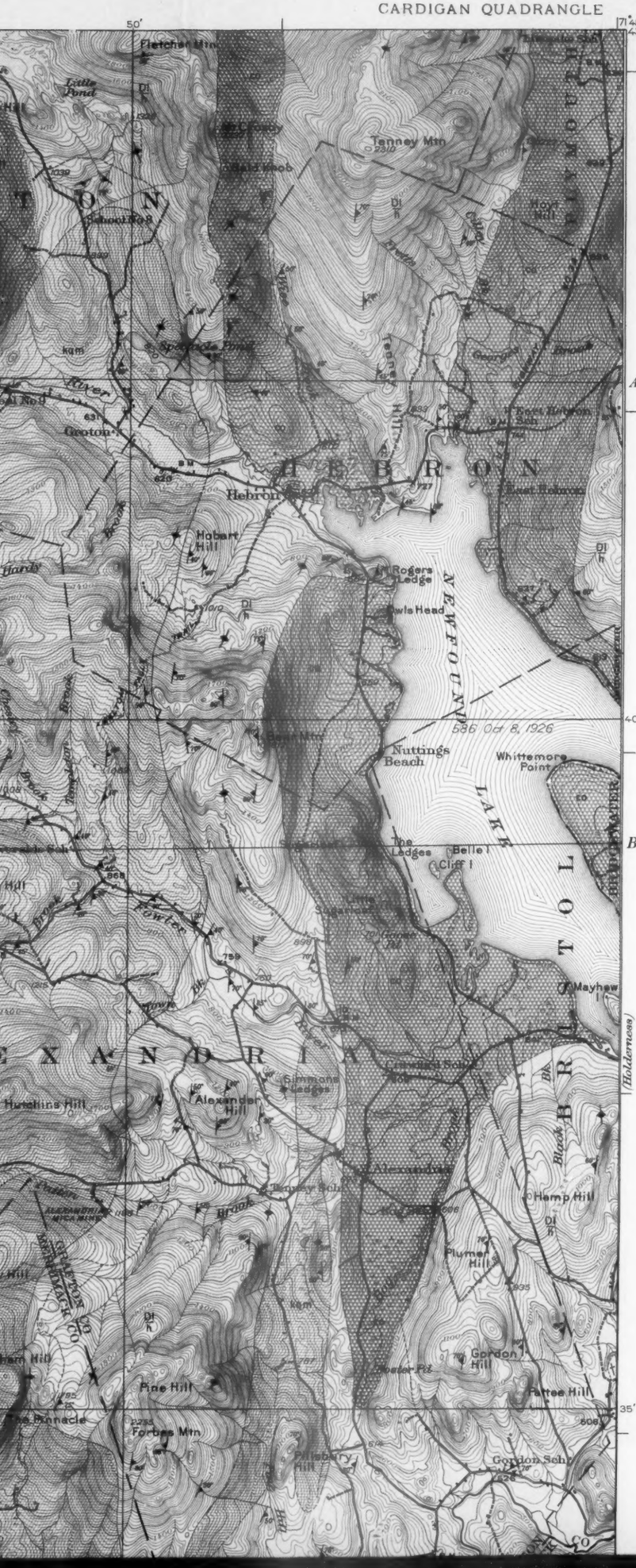
Fowler-Lunn, Katharine, and Kingsley, Louise (1937), *Geology of the Cardigan quadrangle, New Hampshire*, Geol. Soc. Am., Bull., vol. 48, p. 1363-1386.



STATE OF NEW HAMPSHIRE, HIGHWAY DEPARTMENT
FREDERIC E. EVERETT, COMMISSIONER AND STATE ENGINEER

(Rumney)
Thompson Hill
Bucks Corner
DORCHESTER
Poplar Knoll Sch.
GROT
Kimball Hill
Bryant Pd.
Hamlen Hill
Cockerpenny
School No 9
Birch Corner Sch.
Rocky Hill
Mt Tug
Story Sch.
The Pinnacle
Orange
ORANGE
Orange Sch.
Orange Basin
Orange Brook
MT CARDIGAN
AMC Lodge
Bailey Hill
Red Sch.
Mirror Lake
Tuttle Hill
Hoyt Hill
(Mascoma)
Orange Road
East Neighborhood Sch.
Crane Mtn.
Brown Mtn.
ALEX
Tenkenny Pond
Ingalls Mtn.
Gifford Sch.
Barton Mtn.
Melvin Mtn.
Bailey Hill
Tinkham Hill
Wild Meadow Sch.
Shepard Hill
Grants Road
GRAFTON
Tunnicle
Hillsbury Mtn.
East Grafton
Kilburn Pond
Ruggles
Fowler
Brook
Pond
Mine
Creek
Hollow
Road
Sch.
Center
Site
Cem.
Mtn.
Hill
Pond
Brook
Creek

# NEW HAMPSHIRE
# CARDIGAN QUADRANGLE




## LEGEND

### METAMORPHIC AND IGNEOUS ROCKS

| | | | |
|---|---|---|---|
| YOUNGER THAN LOWER DEVONIAN, PROBABLY LATE DEVONIAN | *New Hampshire Magma Series* | *Post-folding* | **Intrusive breccia** (*Angular fragments of schist, quartz monzonite, and granite embedded in a matrix of fine-grained Concord granite; large aggregations of garnet are present; fragments vary in size from a few inches to several feet.*) |
| | | | **Concord granite** (*Fine-to coarse-grained, light-gray granite, weathering brown, composed of potash feldspar, oligoclase, quartz, biotite, and muscovite; locally gneissic.*) |
| | | | **Pegmatite** (*Coarse, interlocking crystals of feldspar, quartz, biotite, and muscovite, ranging in size from a few inches up to several feet in length; feldspar is generally microcline, orthoclase, and rare albite; feldspar crystals 15 feet long have been observed in the Ruggles mine; black tourmaline, garnet, beryl, apatite, montmorillonite, and the rare uranium minerals, as well as rare lithium-manganese phosphates, and metallic sulphides are among the accessories. There are many pegmatites too small to be shown on the map.*) |
| | | *End of folding* | **Kinsman quartz monzonite** (*Medium-to coarse-grained, gray, quartz monzonite, composed of quartz, oligoclase-andesine, potash feldspar, biotite, and muscovite; large phenocrysts of potash feldspar, up to 6 centimeters in length, are a conspicuous feature of the rock, as well as knots and stringers of quartz and feldspar; foliation is generally well developed.*) |
| | | | **Bethlehem gneiss** (*Medium-to coarse-grained, gray granodiorite or quartz monzonite, composed of andesine, quartz, biotite, microcline, and a little muscovite; generally highly foliated; locally contains small phenocrysts or stringers of light-colored minerals.*) |
| | *Oliverian Magma Series* | *Pre-folding* | **Smarts Mountain group** (*Fine-grained, white to gray, weakly foliated quartz diorite and granodiorite; composed of oligoclase, quartz, biotite, hornblende, and microcline.*) |
| DEVONIAN | | *Lower* | **Littleton formation** (*Zone h; green to gray quartzite, mica schist, quartz-mica schist, sillimanite schist, and garnet schist.*) |
| | *UNCONFORMITY* | | |
| PROBABLY ORDOVICIAN | | | **Ammonoosuc volcanics** *Zone m: amphibolite and micaceous quartzite.*) |

### METAMORPHIC ZONES

Shown below formation symbol thus: $\frac{DI}{m}$

m-middle grade

h-high grade

### CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

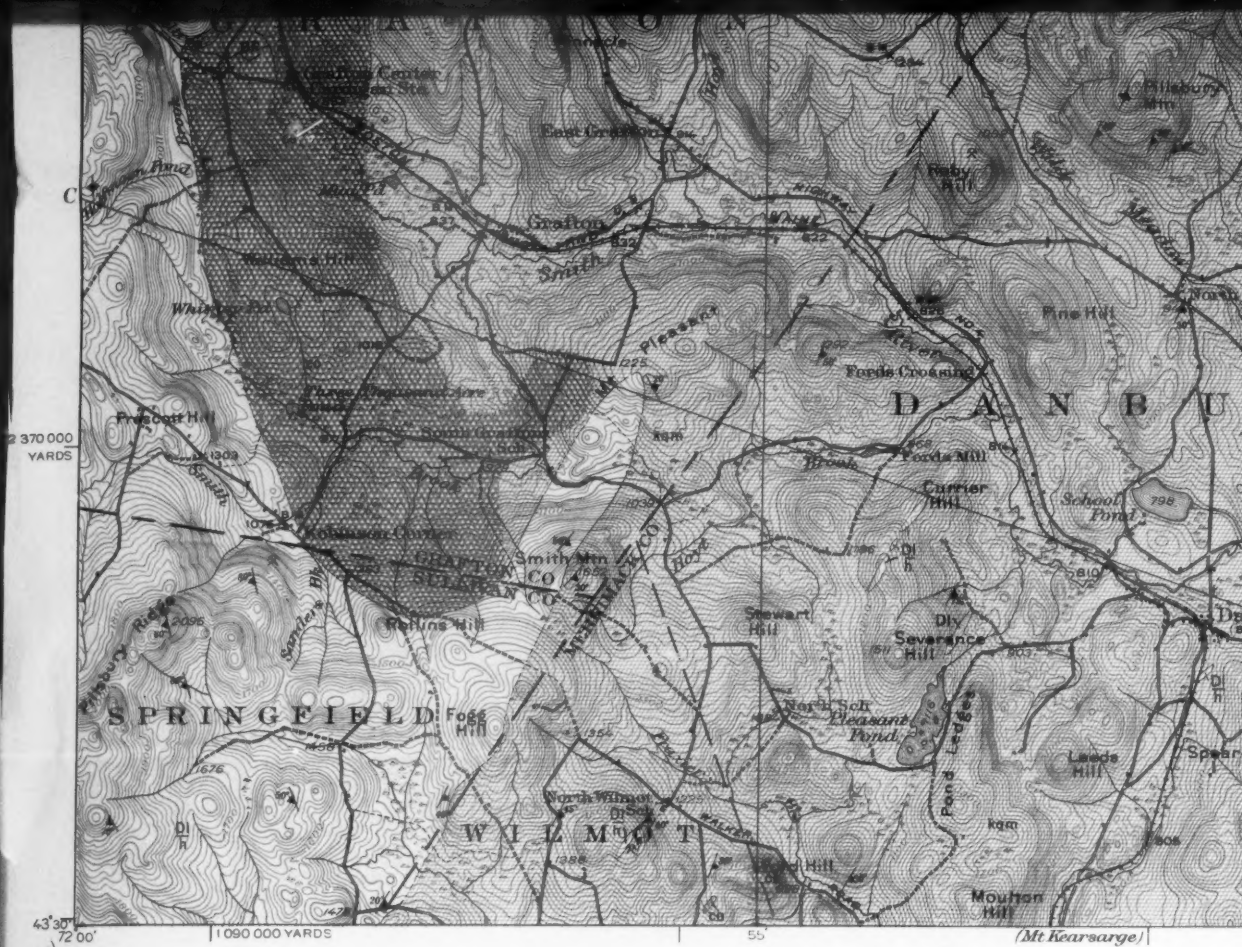
Indefinite as sharp contact is lacking

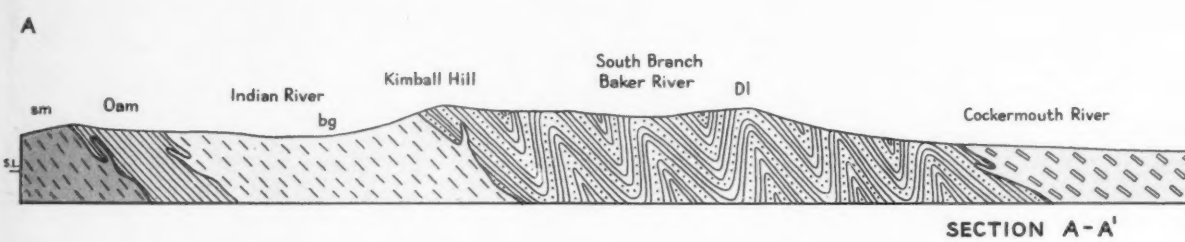
### SPECIAL SYMBOLS

(*Dip and strike symbols representing only a small percentage of the field observations*)

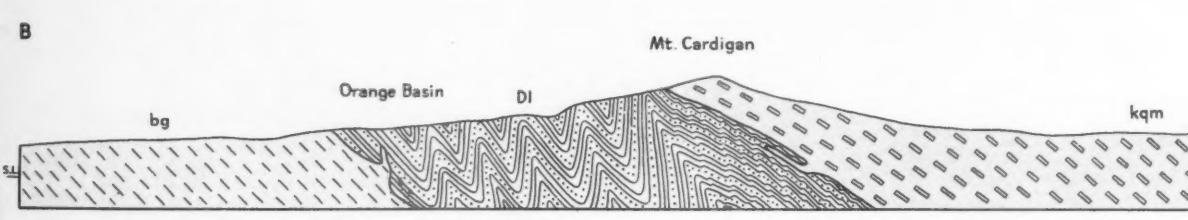
75° — Strike and dip of foliation and schistosity. In general, the bedding, if present, is parallel to the schistosity

Strike of vertical foliation and schistosity

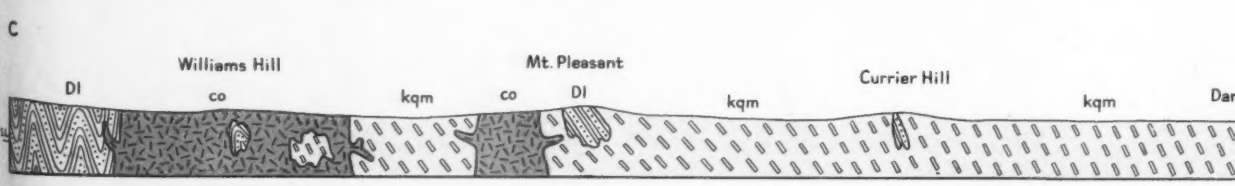



SECTION A-A'




SECTION B-B'

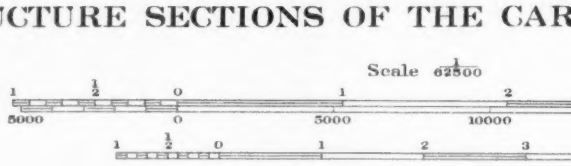



SECTION C-C'

# GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE CAR

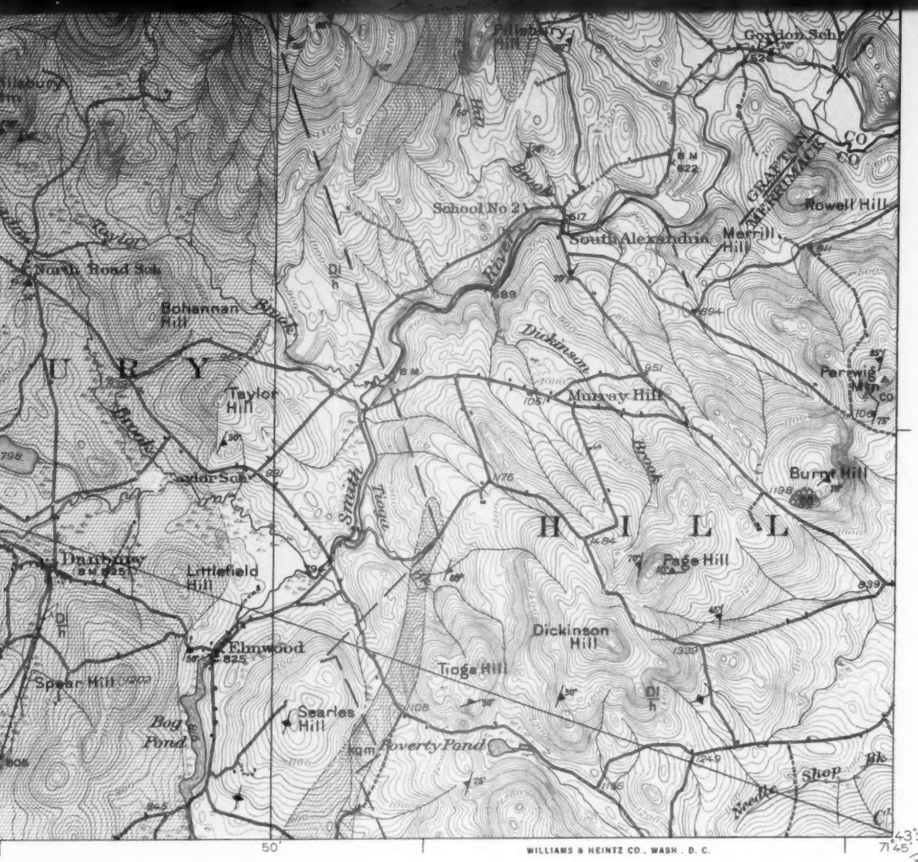
Topographic base by U. S. Geological Survey, surveyed in cooperation with the State of New Hampshire.

Scale 1/62500




Contour interval 20 feet

*Datum is mean sea level*

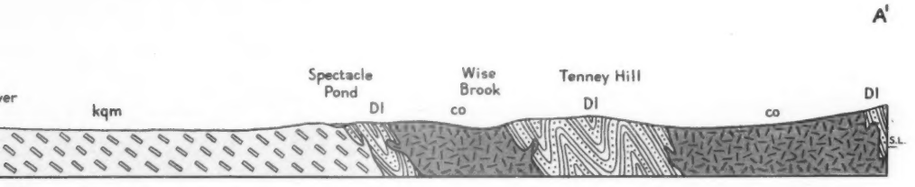


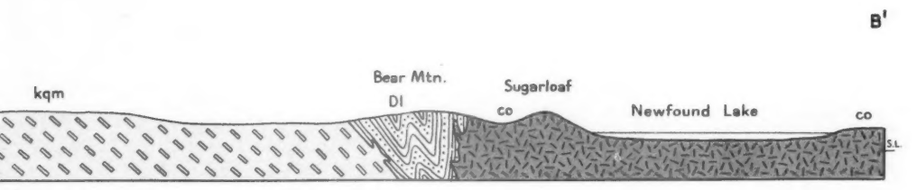

Strike and dip of foliation and schistosity. In general, the bedding, if present, is parallel to the schistosity

Strike of vertical foliation and schistosity

D, Downthrow side of probable fault

Mines, prospects, and quarries, many abandoned



# E CARDIGAN QUADRANGLE, NEW HAMPSHIRE

0 2 3 4 Miles
10000 15000 20000 Feet
3 4 5 Kilometers
20 feet
a level

Geology by Katharine Fowler-Billings and Louise Kingsley
Geology surveyed in 1935 and 1936. Published 1940.
Map edited by Marland Billings.